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ERRATICS AND ARKOSES IN THE MIDDLE PENNSYLVANIAN HAYMOND FORMATION OF THE MARATHON AREA, TRANS-PECOS, TEXAS

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ABSTRACT

Fragments ranging in size up to a maximum of 135 feet in greater dimension, imbedded in an arkosic or greywacke matrix, and derived from a rock succession about 10,000 feet in thickness, are found in several horizons of the Middle Pennsylvanian Haymond formation in the Marathon Basin, Trans-Pecos, Texas. Truly exotic boulders of metamorphic and plutonic rocks and of a mid-Pennsylvanian marine limestone occur with fragments of rocks indigenous to the area. Important folding had occurred by the time the boulder-arkoses were formed; subsequently the Haymond strata were greatly deformed by folding and thrusting. It is thought to be demonstrated that the boulders and their matrix are not thrust mylonites attributable to any of the visible faulting, but that they are really sediments. Possible modes of origin are considered to be either that they were derived as a consequence of possible diastrophic events not now definitely ascertainable or that they are ice-transported.

The occurrence of coarse detritals in the Haymond formation was first announced by P. B. and R. E. King.[1] Baker's unpublished field notes of 1915 record very angular chert fragments varying from $\frac{1}{32}$ inch up to 2 feet or more in size in a matrix of greenish-brown sandstone at a locality in the southeastern part of the Marathon Basin east of the Dimple exposure and near the Comanchean contact south of Maxon section house. This locality has not been studied further. The most striking exposures yet known lie northwest of and near the axis of a large syncline east of Haymond station near

[1] P. B. King and R. E. King, "The Pennsylvanian and Permian Stratigraphy of the Glass Mountains," *Univ. Tex. Bull. 2801* (1928), p. 114; P. B. King, "Geology of the Glass Mountains," Part I, "Descriptive Geology," *Univ. Tex. Bull. 3038* (1931).

the eastern margin of the Marathon Basin. The southwestern end of these outcrops was discovered by P. B. King, of the U.S. Geological Survey, accompanied by A. G. Nance and John Bean of the Gulf Production Company in November, 1930. The writer was guided to the locality by Dr. King, and the nature of the deposit was such that Dr. E. H. Sellards and the writer undertook a study of the problems it presented. The outcrops were traced a distance of 8 miles, and the exposures found earlier by the Kings 4 to 7 miles southeast of Gap Tank in the northeast part of the basin were restudied. Much of the data herein presented were obtained by field work done jointly by Dr. Sellards and the writer, although King continued the study of the exposures during the summer of 1931. Much credit is due to both Dr. Sellards and Dr. King for the great amount of aid and information they have given the writer.[1] A number of other geologists have subsequently visited the exposures. The writer has had the benefit of the views of some of these but is alone responsible for the opinions here expressed.

Since Dr. P. B. King's report on the geology of the Marathon area, prepared after many months of painstakingly accurate field investigations, is soon to be published by the U.S. Geological Survey, the general geology is not discussed here except in so far as it bears on the nature and the origin of these particular detritals. It is hoped that what is presented here will prove to be entirely without prejudice to Dr. King's views and to the credit to which he is fully entitled for a truly masterly research in a region of exceptionally difficult structure and stratigraphy.

HAYMOND AND ASSOCIATED FORMATIONS

The pre-Permian Carboniferous of the Marathon region, composed of a prevailingly clastic succession as much as 12,000 feet thick, has been divided in upward succession into the Tesnus, Dimple, Haymond, and Gaptank formations. The three lower forma-

[1] Papers dealing with this problem already published or announced are: P. B. King, C. L. Baker, and E. H. Sellards, "Erratics of Large Size in the Carboniferous of Texas," abstract, *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), p. 200; E. H. Sellards, "Erratics in the Pennsylvanian of Texas," *Univ. Tex. Bull. 3101* (1931), pp. 9–17; P. B. King, "Large Boulders in the Haymond Formation of West Texas," preliminary abstract printed. Paper presented at Tulsa meeting of Geol. Soc. Amer., December 29–31, 1931.

tions, overlying the Devonian (?) with erosional unconformity, are conformable among themselves and are all folded greatly by late Pennsylvanian diastrophism. Their total maximum thickness is about 10,000 feet. The boulder horizon is near the top of the section.

The Tesnus formation, at the base of the series, consists of a possible maximum of 2,000 feet of predominant shales surmounted by 5,000 feet of predominant sandstones. The middle member of the group, the Dimple, is 300–1,100 feet thick, and is chiefly dark limestone in moderately thick beds, with some interbedded shale and chert and a few conglomeratic beds. The Haymond above is alternating sandstones and shales. The Dimple formation contains a sparse marine fauna, chiefly of brachiopods, some of which have been described by Girty in King's recent publication.[1] More recently King has found goniatites in the formation. One thin limestone layer high up in the Haymond contains marine Fusilinella and a small brachiopod, but for the most part the only indications of fossils in this and the Tesnus formation are the fragmentary remains of land plants. The plants of the Haymond are considered tentatively to be of Pottsville age by David White.[2]

The Tesnus and the Haymond formations are very similar in mode of origin, lithology, and color. Both are composed mainly of alternating shales and sandstones, the predominant color of which is dark olive-green. Some of the shales are dark blue-gray to black, and, in a few places in the Tesnus, chocolate. Some of the sandstones are brownish and whitish. Ripple marks are common on the surfaces of the sandstones. The sandstones and shales alternate in layers varying in thickness from a fraction of an inch up to many feet. Bedding is generally thinner in the Haymond and, in fact, through hundreds of feet of section consists of alternations a fraction of an inch in thickness of carbonaceous shales and of thicker green sandy shale or sandstone (Fig. 1). According to King,[3] Dr. David White, after a field visit, pointed out the close resemblance of this stratifica-

[1] *Ibid.*, p. 39.

[2] P. B. King, "The Geology of the Glass Mountains," Part I, *Univ. Tex. Bull. 3038* (1930), p. 43.

[3] Personal communication.

tion to typical varve structure. Dr. Hans Stille has remarked to the writer on the close resemblance between the Tesnus and Haymond formations and the Flysch of the Alps and Carpathians.

The total thickness of the Haymond formation in the boulder-bearing syncline east of Haymond station approaches 2,500 feet, but its top is here and elsewhere eroded, and the sequence ends some distance above the boulder beds. Farther to the north, near Gap Tank,

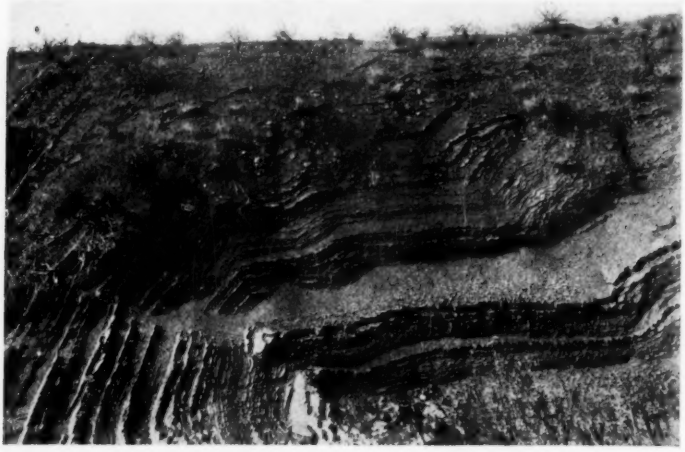

FIG. 1.—Varve-like interbeds of shale and sandstone or sandy shale, Haymond formation, horizon below boulder-arkose strata, exposures on San Francisco Creek, northwest of boulder-arkose locality in syncline southeast of Haymond.

the suspected northern extension of the boulder horizon is separated by 400 feet of shales and sandstones from the top of the formation. The lower part of the succession near Haymond station consists of thinly bedded strata of the sort just described, with a few thin layers of coarser sandstone. Higher up are thick arkose beds, which are succeeded in turn by boulder-bearing strata. The latter are overlain by shales with some sandstone.

The mode of deposition of the Tesnus, Dimple, and Haymond is in doubt. Certain resemblances between these formations and relatively recent deposits of the Colorado River in California have been observed by the writer. On the west side of the Salton Sink—in the

drainage basin of Carrizo Creek—there is a thick series of gently and broadly folded incoherent sands and silts which alternate in thin bands. The silts are flesh colored to orange-chocolate, the sands are lighter hued. At widely separated intervals there are thin reef deposits of the shells of *Ostrea*, *Anomia*, and *Pecten*, relics of brief marine or brackish-water incursions. The writer has interpreted the strata as older delta deposits of the Colorado River, the thicker sand beds having been deposited during the river's flood stages and the silts during the river's low-water stages. A close analogy can perhaps be drawn between them and the Tesnus and Haymond, the latter two being likely of similar deltaic origin.

Both the great thickness and the predominant fineness of grain of the Tesnus, Dimple, and Haymond indicate their derivation from a large or high land mass situated at a considerable distance from the depositional site. The closely folded Paleozoic rocks are covered on all sides by later sediments, but the known extent of Tesnus outcrops covers an area of 800 square miles. The formation has been found in borings beyond the limits of outcrop, but its full extent underlying the later formations is undetermined. The land mass, which was the source of the sediments, lay most probably to the south or southeast of the present outcrops.

THE EXPOSURES EAST AND NORTHEAST OF HAYMOND STATION

The exposures east of Haymond station may be considered as the typical locality, since there the very large erratics are most abundant and the matrix and boulder-bearing horizons are well exposed. At this locality the outcrop extends from $2\frac{1}{2}$ miles south of the Southern Pacific Railroad northeastward for a distance of 8 miles, or to within $\frac{1}{2}$ mile of the Marathon-Sanderson highway.

The deposit gradually wedges out at a point $2\frac{1}{2}$ miles southwest of the railroad up the pitch of the syncline. From its southwestern limit to at least 1 mile northeast of the railroad the boulder-bearing strata are paralleled on the northwest by the outcrop of an underlying thick bed of coarse olive-green arkosic sandstone, but the two are separated by an interval of sandstone and shale. One and two-tenths miles southwest of the end of the outcrop of the boulder bed the arkose beds turn around the axial trough of the overturned syn-

cline, the plunge of which is there gently to the northeast. Southwest of the railroad and for some undetermined distance to the northeast only one boulder horizon is visible, but in the section 1 mile west of the south peak of Housetop Mountain—called House Mountain on the Longfellow Reconnaissance Topographic Quadrangle—six beds are exposed in the section. The width of the outcrop here—at right angles to the strike—is at least ¼ mile, and the dip of the strata is approximately 45°. The boulder beds are separated by sandstone and shale. Dr. King finds that at one locality the boulder beds extend through 900 feet of strata and that single lenticular boulder-arkose beds are up to 150 feet thick.[1]

The following additional data are quoted from Dr. King's letter:

> The boulder bed is not faulted off to the north, but continues to within one-half mile of the Marathon-Sanderson road where it passes under thick terrace gravels. The big thrust fault to the east runs under the corner of Housetop Mountain. The shales east of the boulder bed are an upper part of the Haymond and not Rough Creek (the lower member of the Tesnus). The axis of the syncline is in them. The more exposures are seen the more remarkable does the matrix of the bed appear. It is an arkosic sandstone with few bedding planes. Its contrast with the typical slabby Haymond is great. The boulder beds are long lenses. Two of them persist from Housetop Mountain to near the Sanderson road, but at the Housetop Mountain locality two more beds wedge in above and two below. The big Pennsylvanian limestone boulders are *clustered*. For a stretch of a mile or so none larger than two-foot masses will be seen. Then for a mile or so they will be so thick they are hard to count—then they disappear again. There are five such big clusters between the Sanderson road and the Bates place. The Pennsylvanian fossiliferous limestone boulders range in size from huge boulders to minute pebbles. The big ones are not at the base of the series but in all six members, but most abundantly in the middle two.

Matrix of the boulder beds.—The nature of the matrix is shown in Figures 2 and 3, a more consolidated phase of which is shown in Figure 3. The finer matrix comprises on the average perhaps $\frac{9}{10}$ of the total exposures. This matrix is mostly unstratified and usually loosely coherent. It and the boulders it contains are totally unassorted. Partly because it is relatively easily eroded and partly because of the extensive débris covering the middle of the known outcrop, it is well exposed only locally. The matrix proper consists of a

[1] Abstract of paper to be presented before the Geological Society of America, Tulsa meeting, 1931.

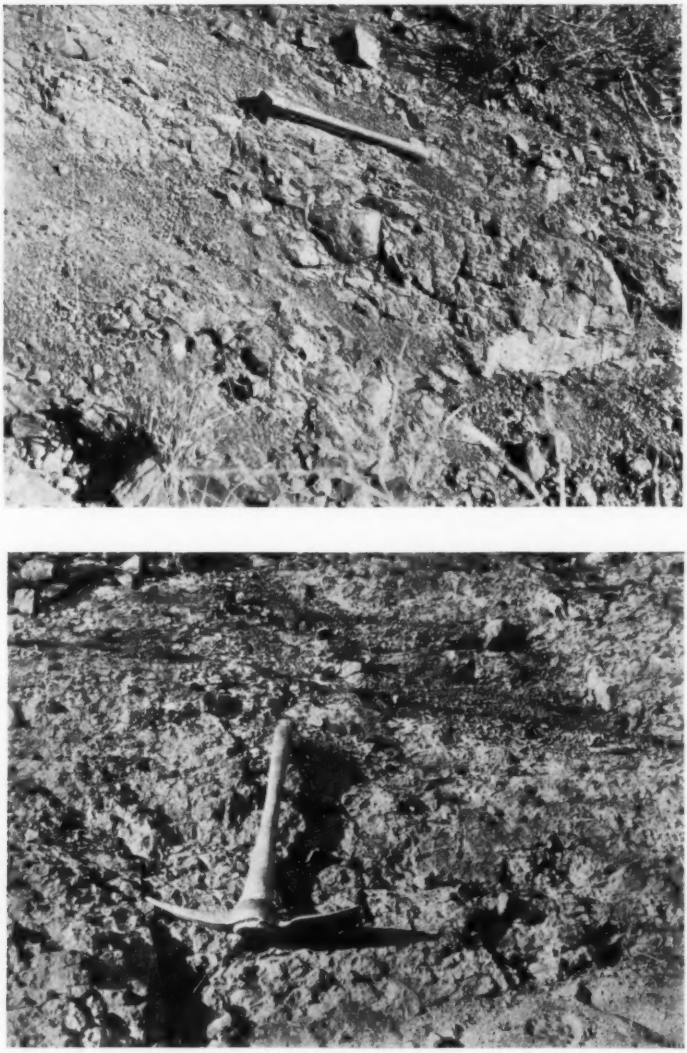

FIG. 2.—Less consolidated phase of boulder-arkose beds, showing characteristic matrix and small sparse embedded boulders. Locality west of Housetop Mountain.

A

B

FIG. 3.—More consolidated phase of boulder-arkose: *A*, west of Housetop Mountain; *B*, southeast of Haymond station.

fairly coarse, gritty, olive-green arkose or graywacke of angular particles of quartz, feldspar, mica, chert, and Tesnus sandstone with a few worn crinoid stems and broken *Archaeocidaris* spines. Characteristically it has small white spots, perhaps of kaolinized feldspar. Pebbles, boulders, and larger erratics, varying in larger dimensions from less than $\frac{1}{10}$ inch up to more than 100 feet, are scattered so sparsely for the most part and with such widely varying sizes throughout the matrix that it is impossible they could have been deposited by the more ordinary aqueous agencies. There are some interbeds, however, of thin, well-bedded, fine-grained sandstones and finely laminated shales.

Mr. George H. Otto, under the direction of Dr. F. J. Pettijohn, of the University of Chicago, has examined hand specimens and thin sections of the matrix rock, and a summary of his findings is as follows:

> The rock might best be called a conglomeratic greywacke or a conglomerate with a greywacke matrix, composed of about 50 per cent mineral grains, 30 per cent rock fragments, and 20 per cent finer matrix. In the mineral grains, quartz and feldspar predominate, and there are minor amounts of chloritized biotite, muscovite, and secondary veinlet and replacement calcite. The quartz grains show all degrees of rounding and some of them have abundant rutile inclusions. The quartz is notably fractured and exhibits strain shadows. Feldspar, which is not greatly altered, is about one-third as abundant as quartz. Biotite, apparently allogenic, appears as worked and crumpled plates. Muscovite may be secondary after biotite, that which is associated with feldspar apparently being secondary. Rock fragments are very abundant and include especially quartzite, chert, slate, limestone, and a smaller number of metamorphic rock fragments. The finer matrix resembles the material of fine-textured shale. The most characteristic features of this rock are the lack of sorting and stratification, the high content of feldspar, the variation in angularity of the mineral grains, the variety and abundance of rock fragments, and the fine-grained matrix. It is clearly not a normal sediment derived by the ordinary processes of rock weathering. Rock disintegration rather than decomposition seems to have been the dominant weathering process. Rapid deposition with consequent lack of sorting and stratification appears evident. The most probable origins suggested are glacial, that is, the deposit is a tillite, or alluvial fan deposition, or, in other words, the deposit is a fanglomerate.[1]

Pebbles, boulders, and erratics.—In the boulder beds pre-Cambrian rocks are present in all sizes up to 1 foot in the larger dimension.

[1] Letter from Dr. F. J. Pettijohn.

They consist of porphyritic and granitoid, mainly acidic plutonics, gneisses, schists, vein quartz, pegmatite, aplite, and a finely conglomeratic gray quartzite. A large pebble of hematite, observed in the beds, may also be pre-Cambrian. Many of these stones are fairly well rounded, but a considerable proportion of them are sub-angular. Some are very conspicuously "soled" or flattened on one or more sides. A number are very highly polished on one or more faces. All the pre-Cambrian rocks, and all, except the vein quartz and quartzite, are decomposed considerably and disintegrated on the outside. Few striae have been found yet on the pre-Cambrian rocks, but one small pebble has cross-striae and one large quartzite pebble is well scored and polished on one face. It would seem quite possible that the pre-Cambrian rocks were transported from original ordinary gravel deposits.

The Paleozoic formations of the Marathon area contain many different cherts of various appearances and colors and include the white chert-like Caballos novaculite. The most abundant of these in the detritals are novaculite, but the black upper Maravillas and green Caballos chert fragments are represented abundantly. The largest silicious erratics are the novaculite which range in size up to at least 30 feet in larger dimension. Some of the novaculite blocks are contorted, others are of reconstructed, greatly brecciated novaculite, with angular brecciated fragments recemented in a matrix of the same material; others are of the original undeformed rock. Many fragments of bedded chert are commonly angular, no matter how worn, and are likely to be scored, fluted, grooved, striated, and highly polished wherever they have suffered intense deformation. Hence, such phenomena are not distinctive of glaciation when exhibited by cherts such as these. For instance, the scorings shown in Figure 4 can be either glacial or slickensides produced by fault movements, but, if fault-slickensided, this particular block, 15 feet in larger dimension, was not eroded subsequently. One large boulder of Maravillas chert is deeply grooved on one side and exhibits numerous cross-striae on one flattened surface. Percussion marks are likewise common on the chert surfaces, as well as on the other rocks. There are blocks of very homogeneous, hard, dense, fine-grained, dark olive-green Tesnus quartzitic sandstone, which exhibit no evi-

FIG. 4.—Slickenside-like markings on surface of boulder

FIG. 5.—View of large exotic blocks of Middle Pennsylvanian (Lower Strawn or Millsap) limestone, west of Housetop Mountain. The largest block shown is 100 feet long.

dence of lamination, and which are mostly coated with a film of brown iron oxide. A number of the blocks have shapes similar to those found in glacial deposits. Some of the Tesnus sandstone blocks are at least 12 feet in larger dimension. The weathering of these rocks by exfoliation has destroyed on many of the boulders any scratches which have been present. The Tesnus boulder, shown in Figure 6, exhibits on one face numerous parallel striae.

FIG. 6.—View of large boulder of Tesnus quartzitic sandstone, near southern end of boulder-arkose exposure southeast of Haymond station. The side facing the observer is marked by numerous parallel striae, running approximately parallel to the surface of the ground. Some of the matrix is adhering to the depressions bordering the smoothed and striated protuberance.

Four limestones are found among the fragments. One is the limestone of the Maravillas chert. Another is Lower Ordovician, recognized by its fossils and thin layers of brownish to blackish chert. These two limestones are indigneous in the area. The other two limestones are of exotic facies (Figs. 5 and 7*a*). The exotic limestone of most of the largest blocks found, which range up to 100 feet in larger dimension, is dark blue-gray, fine-grained, somewhat earthy

A

B

FIG. 7.—*A*, Embedded block of exotic limestone, west of Housetop Mountain. *B*, Angular block of unbrecciated novaculite, measuring 10 by 6 by more than 2½ feet, east end of northern exposure of Haymond conglomerate, 5¼ miles east-southeast of Gap Tank.

in texture, in part siliceous, very massively bedded, and it contains irregular segregations of rust-colored chert and silicified fossils.

Fossils from this limestone were submitted to Raymond C. Moore, Carey Croneis, and F. B. Plummer. Mr. Plummer states that the limestone is lower Strawn Pennsylvanian and has identified the following fossils:

**Productus inflatus* McChesney (most common)
**Composita subtilita* (2) (Hall)
Chonetes geinitzianus Waagen (1)
Squamularia perplexa McChesney (1)
**Spirifer rockymontanus* (?)
Spirifer rockymontanus ? (Marcou) (?)
**Lingula tighti* Herrick ?
Spirifer sp. indet.
*Marginifera sp. ? New variety with small nodes and wrinkles
Derbya sp. indet.
Composita sp. ?
Trachydomia wheeleri (Swallow)
Cordaites sp.
Corals, probably Trilophyllum. The form resembles specimens occurring in the Smithwick and Lower Strawn.
Fusilinids

These fossils are definitely Middle Pennsylvanian. The fauna as a whole resembles that from the Pennsylvanian of the Big Lake oil field, those starred in the foregoing list being found at both places. Dr. Croneis, however, determined the age of the fossils as early Canyon, referring them specifically to Graford time.

Numerous boulders of a very rough-surfaced concretionary, nodular, fossiliferous limestone which do not show any evidence of wear are embedded in the deposits. The large limestone masses have rough, angular surfaces. The larger masses are generally tabular—relatively thin compared to length—but some are broad and flattened on the ends. The larger more tabular masses are embedded parallel to the strike. They are fractured in places, and portions of them are displaced by small dip faults which traverse the strata. Although embedded in the coarse deposits, the limestone blocks contain no detrital material. They are thought, therefore, to be exotic to the formation, this view being strengthened by their occurrence in all six boulder horizons.

There are boulders of Dimple limestone in the boulder bed south of the railroad. One of these measures 135 feet in larger dimension and has been bent into an L-shaped mass by deformation subsequent to its deposition. The enclosing strata are similarly deformed.

FIG. 8.—Typical exposure of Haymond conglomerate, in northern belt of exposures, west of road and 4.4 miles east-southeast of Gap Tank.

THE EXPOSURES SOUTHEAST OF GAP TANK

About 14 miles north of Haymond station is a second locality at which erratics are found in the Haymond formation. At this locality a conglomerate occurs in the Haymond in a horizon 1,400 feet above the base and 400 feet below the top (Fig. 8). The con-

glomerate horizon can be followed with some interruptions for about 4 miles. The exposures are at the northeast side of the Marathon Basin close to the Cretaceous rim, and at places the conglomerate is covered by Cretaceous. This conglomerate was mapped by Philip B. and Robert E. King and is shown on their geologic map of the Glass Mountains.[1]

The conglomerate at this place appears to average about 10 feet in thickness. It is associated with sandstones in part coarser-grained than the Haymond sandstone in the first-described area. Some of the sandstones are cross-bedded and contain shales interbedded with them. The conglomerate grades upward and downward for a short interval into conglomeratic sandstones containing a few angular to subangular small pebbles. The conglomerate consists of closely packed, unassorted fragments, mainly novaculite with subordinate amounts of Upper Ordovician cherty limestone. The conglomerate contains Tesnus sandstone, one piece 2 or 3 feet across having been observed. A small boulder of porphyry was also found. There are a number of large novaculite boulders, one measuring 10 by 6 by more than $2\frac{1}{2}$ feet. Near the one figured, there was a large angular block of fossiliferous Ordovician limestone. The conglomerate is remarkable in the following respects:

1. The lack of grading in size or assortment.
2. The very small amount of finer matrix in it.
3. Its thinness.
4. Its persistency.
5. It is the only conglomerate in the section.
6. Many of its novaculite boulders show brecciation and recementation with novaculite before transportation.
7. The general angularity of the fragments, although novaculite fragments are never well rounded, the brittleness of the rock forming sharp angular contacts between faces. The limestones are, however, rounded only when exposed at the surface where they have undergone solution-etching since the conglomerate bed has been exposed. In such limestone fragments, the cherty parts project above the surfaces of the weathered limestone.

[1] P. B. King, "Geology of the Glass Mountains, Texas," Part I, "Descriptive Geology," *Univ. Tex. Bull. 3038* (1931).

DEFORMATIONAL HISTORY

We find in this area great structural complication and long-continued orogenic movements. There is clear evidence of at least three mountain-making disturbances occurring in the time-interval between the beginning of Tesnus deposition and the end of the final folding and thrusting of the Haymond and older formations. Either there was gradually increasing intensity of movement until a maximum at the time of the post-Haymond overfolding and overthrusting or the zone of most intense deformation moved farther to the northwest in later time.

The Tesnus formation is unconformable upon the Caballos novaculite and chert. The Tesnus is a terrigenous deposit, with considerable greywacke and arkose. The proportion of greywacke and arkose increases and the formation apparently thickens toward the southeast. There are no arkoses in the earlier, post-Cambrian sediments of the area. Therefore, the Tesnus sediments came from a new source, a land area newly uplifted and presumably rapidly eroded. Hence the diastrophism appears to have begun as early as Lower Tesnus, although the source of the sediments was probably then at a considerable distance, as conglomerates are rare in the Tesnus, are fine in texture, and are mainly composed of detritals from the earlier sedimentary rocks.

A second epoch of diastrophism, or the extension of diastrophism farther northwest, is demonstrated by the boulder deposits of the Haymond. This diastrophism was intense, since the novaculite was greatly deformed before boulders derived from it were deposited in the Haymond, and there must have been land of high relief to furnish the extremely coarse detritals from the thick succession of rocks which are found in the Haymond boulder beds.[1]

A third deformative epoch, or a continuation or extension of the older movements, is the one during which the Haymond formation was strongly deformed. Important thrusting movements occurred either during this third phase or somewhat later.[2]

[1] We assume here what we will afterward prove, namely, that the Haymond boulder beds are sediments.

[2] This third phase may be divided ultimately into two or three epochs of renewed intensity. We will merely state without presenting evidence that there may have been deformation in (1) post-Haymond and pre-Gaptank time; (2) post-Gaptank and pre-

The overthrusts belong to the third phase. There is a possibility that thrusting may have occurred during the second, or Haymond, epoch, but there is no proof and the only evidence is the deposits which are being considered. Some of the overthrust planes are folded, but such folding may have been essentially contemporaneous with a single epoch of thrusting. Also, there may be two epochs of thrusting, with folding of the older thrust planes during the second.

EROSIONAL HISTORY

It is certain that Haymond and older strata have been greatly eroded in at least three later epochs: (1) post-Haymond and pre-Hess; (2) post-Permian and pre-Cretaceous; (3) post-Cretaceous. The writer thinks there was another erosion interval, in later Pennsylvanian, after the Haymond and before the Gaptank deposition, but this is doubted by P. B. and R. E. King, Schuchert, Blanchard, Keyte, and Baldwin. Possibly, also, the strata were eroded in an interval between Gaptank and Wolfcamp deposition, and during part of Permian time as well.

POSSIBLE MODES OF ORIGIN OF THE BOULDER-ARKOSE

It is conceivable that the Haymond boulder-arkose might be attributed to one of the three following modes of origin:

1. Mylonites (thrust or "normal" fault breccias)
2. Sediments transported and deposited by agencies other than ice
3. Ice-transported sediments

Efforts to interpret the phenomena described in the preceding sections are handicapped because of the very limited exposure of the boulder-arkose beds and the overlap of Cretaceous strata which covers the territory to the north, east, and south. The erratics contain rocks from a succession of strata in the immediate area found to have a thickness of 10,000 feet or more. This fact must be kept in mind in any attempt to explain their occurrence.

It is perhaps to be recognized at the outset that the occurrence of

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Wolfcamp time; and (3) post-Wolfcamp and pre-Hess time. The evidence consists both in the basal conglomerates and suspected foldings (in places the Wolfcamp has greater dips than the overlying Hess, and some of its upper strata are missing).

scratched, grooved, soled, and polished pebbles and boulders in strata which have been compacted and strongly tilted during severe deformation has very likely no particular significance. Such phenomena can be caused through interstitial movements grinding together, under great pressure, the pebbles and boulders. However, some of the soled boulders have been shaped more likely by the agency of ice.

Although the soling is not restricted to any particular rocks it is shown best in quartzites, mostly of probable pre-Cambrian age, in quartzitic sandstones of the Tesnus formation, and in the various cherts. These are the rocks least affected by subsequent alteration.

The soled boulders have at least one conspicuously flattened surface. One has seven flattened surfaces, two of which are flattened more than the others. The soled surfaces are not entirely plane in all boulders. Those of larger size are concave in many cases, some slightly, others markedly so. Those less completely soled are rougher and more irregular. The edges between the soled surfaces of the quartzites are generally rounded off, but in some of the cherts the edges are sharp. The soled surface of the gritty textured quartzite of another boulder bevels nearly at right angles both the lamination planes and a system of closely spaced joint planes, which cut the lamination planes at an angle of about 30°. None of the five soled surfaces on this boulder are parallel to either the lamination or the jointing. A boulder of homogeneous (non-laminated) Tesnus sandstone has four conspicuously flattened surfaces converging toward a somewhat pointed, but rounded, end. Although this rock is not sufficiently indurated to preserve striae well, traces of striae are still preserved on all four faces. The narrowest and least soled face of the four shows plucking of material from a weaker, fractured band.

It is quite evident that these soled boulders have not been shaped in the main during transport by running water, although some of the edges between the flattened surfaces may have been worn off by the more ordinary processes of attrition. Because these boulders have long been exposed on the present surface, some of their edges quite likely have been rounded by exfoliation, which is always most effective on sharp edges. In this case it is particularly effective, for the edges no doubt were weakened, strained, and fractured by the forces

which did the soling. Exfoliation since exposure is proved to be an important factor, as many surface boulders are prominently fractured, and in many cases some part of one already has fallen away from the main mass.

The soling of these boulders has occurred while they were held firmly in some medium which temporarily prevented movement within the medium. At times some of them must have been turned or partly rotated and again temporarily held firmly while another face was soled. Two adjoining faces could be soled simultaneously in a groove along which they were moved. Or single faces of two boulders may have been soled at the same time by the one moving over the other. Such soling may be accomplished by faulting or interstitial movement in the rocks as well as while they are being held firmly in a moving mass of ice. It is doubtful whether there is anything essentially different in the effects which may be produced by either cause.

1. *Mylonites.*—The mode of origin first to be considered in this highly deformed area is the tectonic. Such an origin would explain the erratics and arkose as thrust breccias. But this explanation will not serve for the conglomerate exposed in the northern area, southeast of Gap Tank. The conglomerate dips at high angles in its western part and has good outcrops on steep, bare slopes. Faulting, if it occurred there, would be detected with ease. There is also gradation from the peculiar conglomerate through thin zones of finer and sparser detritals into both underlying and overlying strata.

The strata in the syncline east and northeast of Haymond are not fully and continuously exposed. Although the country is treeless, its surface is largely covered with grasses and dwarf thorn-scrub. A broad alluvium-covered valley in the southwestern part of the syncline follows the strike of the syncline for most of its course. About $\frac{1}{4}$ mile south of the disappearance of the boulder-arkose zone the valley turns south toward its water gap through the ridge of the Dimple formation which forms the southeast rim of the synclinal valley. This water gap is situated some distance northeast of the southwest end of the encircling hogback of the Dimple formation. Just south of the water gap the Dimple strata are vertical, and to the

northeast of it they are overturned and dip steeply (approximately 70°) to the southeast. Still farther northeast the Dimple strata are cut out by the diagonal overlap of an overthrust.

The boulder arkose zone north of the railroad is partly hidden by piedmont gravel deposited on a gently sloping rock-pediment surface beveling the steeply dipping bedrocks. Subsequent dissection of the gravels has exposed in part the underlying bedrock.

The trace of the overthrust already noted intersects at a slight angle the dip of the strata in the syncline, gradually overlapping higher strata to the northeast to approximately the line of the railroad. To the northeast of the railroad the overthrust trace approximately parallels the strike of the underlying strata and at no place in the area of the boulder-arkose outcrop does it cross over the axis of the syncline, but everywhere rests on its southeast flank. The plane of the overthrust dips steeply southeast; it has a stratigraphic displacement of probably 5,000 feet as a maximum; the overriding sheet is Tesnus formation and the "sole" is Dimple or Haymond or, farther southwest, Upper Tesnus, depending upon the amount of displacement in the various sections. Where the overthrust reaches nearest to the boulder zone its trace is obscure because of the general lithologic similarity of the Tesnus and Haymond rocks on the two sides. However, close work serves generally to detect the outcrop of the overthrust by the shattering of the beds in its zone and the difference in strike on the two sides. Some narrow lenses of Dimple are found on the sole of the overthrust in the area near and both south and north of the railroad. Possibly these are not entirely in their proper places in the section, but at any event they are not far removed from it. No blocks or smaller fragments of any older rocks are found along the fault.

The problem is therefore apparently restricted to the possible occurrence of a lower overthrust sheet which has stripped off, more likely from its sole, fragments of all sizes from a rock section 10,000 feet in thickness and carried them forward along the overthrust to their present situation. This is admitted to be possible under conditions of such acute deformation as this area has undergone—and more particularly if the overthrust plane itself had been folded into

a syncline before the time of the overthrust just described. But the evidence, as set forth below, does not favor such a view, and the possible fault has been carefully searched for.

The matrix in which the various erratics are embedded in the exposures of the syncline east of Haymond comprises by far the greater bulk of the assemblage. The boulder horizons range to a thickness of considerably more than 100 feet, are everywhere present along the exposures, are interbedded with soft clay shales; and their matrix is very friable and for the most part practically unconsolidated. The fault-breccias of the area are, on the other hand, mostly firmly cemented either by chert or calcite. Under the pressures necessarily involved in overthrusting, it would seem that the incompetent matrix would have been mostly stripped from the fault plane and squeezed into a thickened mass which would be likely to be segregated in places apart from the boulders, but this is largely conjectural and perhaps not necessarily true. However, we do not find such matrices nor such a varied assemblage of fragments of so many different kinds of rocks along any of the known faults in the region, although some of them do contain horse-like inclusions of the older rocks.

Furthermore, Dr. King's work[1] shows that in both areas the boulder arkose beds are succeeded by higher strata of the same formation, the Haymond. Hence, if a fault existed in the syncline east of Haymond, it must be a mere intraformational sliding within the Haymond itself which could scarcely drag in such an assemblage of erratics, and in any event the fault would have had to be folded into a syncline before the time of the overthrust which is known or as a consequence of its action. In addition, such a supposed overthrust must have had six imbricate sheets or slices—one for each horizon of erratics. The possibility is not entirely disproved, but the probability seems to be against such a fault.

2. *Sediments transported and deposited by agencies other than ice.*—The facts already given make it highly probable, if, indeed, they do not satisfactorily prove, that the boulders and arkoses of the southern outcrops are sedimentary. The conglomerate exposed southeast of Gap Tank is surely sedimentary.

If these erratics, boulders, and arkoses were transported and de-

[1] Geologic map and subsequent, still unpublished, work.

posited by an agency or agencies other than that of ice, the following requirements must be explained satisfactorily:

1. Probable subaqueous or deltaic deposition, at least of the directly underlying and overlying strata.

2. Lack of assortment and bedding of the boulders and matrix.

3. The great preponderance of matrix, compared with the relative sparseness of boulders, pebbles, and huge erratics, in the southern exposures.

4. The interbedding of fine-textured and thinly laminated deposits with the coarse débris.

5. The sudden beginning and ending of the deposition of very coarse detritals, which do not gradually grade off into finer sediments.

6. The "soling" of a considerable number of the boulders.

7. The derivation of (*a*) the coarse detritals from a succession of strata approximately 10,000 feet in thickness and including all the more consolidated formations of the area, and (*b*) the largest boulders from a truly exotic formation.

8. An agency capable of transporting boulders in sizes up to 135 feet in largest dimension.

9. The existence, in the underlying part of the same formation, of a thick succession of what appears to be varved deposits.

Part of the conditions would be fulfilled by débris or talus fans the detritus of which was deposited and transported by being rolled and washed down slopes, or through the agency of mud flows, solifluction, talus creep, landslides, or earthquake slides, or by the combined agencies of two or more of these. Of these agencies mud flows are probably capable to transport farthest really large blocks. Repeated mud flows, moving the blocks in each flow, would transport them probably farthest in the end. Cloudbursts transport large débris farther and more of it than do ordinary rains. Cloudbursts, mud flows, and landslides occur sporadically, and their deposits may be separated by fine-textured sediments. Boulders are gradually shifted down steep-gradient stream courses by running water undermining their supports when it carries away underlying and adjacent fine material.

Coarse débris can be distributed over a wide extent of territory

by an overthrust sheet. Obviously, the front of such a sheet advances for a time at a rate greater than it is worn back by denuding processes. Overthrust sheets in Western Persia are stated to be advancing now at a distinctly visible rate.[1] Débris is yielded by an overthrust frontal scarp both during its forward movement and during its recession, after forward movement has ceased or its slowing rate is overmatched by the processes of erosion.

The coarse débris in the Haymond may have been distributed as waste from a single overthrust sheet or "nappe," the coarser deposits formerly connecting the two remaining outcrops having been destroyed by the subsequent erosion; a sufficiently broad and high anticlinal arch may have been upraised in the area between the two belts of exposures; an overthrust ridge may have existed between them; the two belts of detritals may have been derived from separate overthrust sheets; one overthrust sheet cut in two by erosion may have yielded débris to the two belts; or less probably, as we know of no important "normal" faulting during the late Paleozoic movements, the débris may have been furnished by "normal" fault blocks.

The interbedded fine-textured shales in the boulder-arkoses of the southern belt may have been deposited in intervals between times of repeated advances of an overthrust sheet.

At the present time the nearest exposures of pre-Carboniferous rocks are about 5 miles distant from the nearest part of the boulder beds, and probably these were not exposed until long after the folding of the boulder beds. At any rate, they were still farther away at the time the boulder beds were deposited, because the crustal shortening in the subsequent folding is considerable.

As a matter of fact, there does not remain anywhere, in the part of the area visible, a single vestige of any of the sources conceived in the above. Nor are we certain that any thrusting or "normal" faulting occurred within Haymond time in any visible part of the area. But, conceivably, the "roots" of the northern of two overthrust sheets or the greatly eroded remnants of a former arch or "normal" fault block may be buried in the area between the two boulder-arkose belts. The "roots" of a possible overthrust or over-

[1] H. G. Busk, *Geologic Structures* (England: Cambridge University Press, 1929).

thrusts may be concealed beneath the Cretaceous cover on the southeast, which now reaches to within 10 miles of the southern boulder-arkose belt. Less probably, because all known overthrusts came from the southeast, vestiges of the original source or sources may be concealed under Cretaceous deposits to the east of the southern belt or under Permian or Cretaceous north or east of the northern belt of boulder-arkoses.

The hypothesis, however, encounters other difficulties, besides the lack of any definite evidence supporting it. The high, steep, and nearby slopes necessary for it would be expected to give a greater proportion of coarse débris than we find associated with the arkosic matrix and not to supply the interbedded fine sediments. There should be much more of a gradation in size of the débris, especially into the lower and higher beds. The northern thin but very coarsely fragmental boulder bed must have been derived from the different formations in a rock section about a mile in thickness, even if the Ordovician to Tesnus sediments there had thinned to only half their thickness farther south. Why should the only bed in the section be so thin and yet contain such large blocks? Can the hypothesis explain the varve-like banding of so much of the Haymond deposits? Have the "soled" boulders a satisfactory tectonic explanation? Is not the hypothesis more of a possibility than a probability? Even if it is accepted it does not disprove the possibility of ice-transport.

Ice-transported sediments.—The probability is great that the presence of a mountainous relief must be assumed in order to account for the boulders in any case. High mountains are apt to have glaciers. The writer is not greatly concerned with the essential contemporaneity of these deposits with coal beds in adjoining areas; coal beds in the Southern Hemisphere and in Peninsular India are closely associated with glacial deposits, and in Southwestern New Zealand and Southern Chile today luxuriant vegetation flourishes within a mile of the glaciers. In Southeastern Alaska, coastal British Columbia, and northern Washington state, glaciers, dense forests, and peat bogs are all visible from the same observation point.

The hypothesis of ice rafting appears to be a possible mode of deposition of the Haymond conglomerate exposed southeast of Gap Tank. The largest boulders, there, are of a size so great that aqueous

transportation is very unlikely. The conglomerate was deposited by some agency which began and ended its action suddenly. If water transported all of the boulders and pebbles, they should exhibit more evidence of water erosion, and probably also some signs of assortment. If dropped from blocks of ice either grounded or floating in shallow water, the unassorted fragments, if not later retransported by aqueous agency, would remain in the heterogenous assemblage they exhibit. The absence of a fine matrix can perhaps be explained in either of two ways. The fragments could either have been plucked from the sides of an ice-filled valley or else have fallen from cliffs or rolled down steep slopes onto a surface of ice. Or, if the land areas furnishing Haymond sediment were either covered with ice or frozen at least for most of the time during which the conglomerate was deposited, the shallow water would at that time be free from a load of fine sediment, permitting waves and currents to wash out from the boulder bed the finer particles. Sedimentation of fine-grained detritus, interrupted during the formation of the conglomerate, was resumed when conglomerate deposition ceased.

The hypothesis of ice rafting when applied to the southern exposures encounters some possible serious objections. In these there is no evidence of sorting or stratification, although overlying, underlying, and interbedded strata are extremely well stratified and fine textured. The matrix of the boulders and pebbles, although differing from the associated sediments in being both coarser and arkosic, predominates greatly over the heterogeneous boulders, pebbles, and larger erratics which it encloses. Because of these facts, it is difficult to believe that floating ice alone could have formed such a thick and extensive deposit free from assortment and bedding or lamination. Icebergs grounded on the bottom and shedding their detritus on melting would be a possible explanation, but the deposits are too large. They look much more like moraines.

In conclusion, it may be stated that no evidence discovered so far is opposed to the hypothesis of ice transportation. In fact, every criterion for such an origin except a striated rock pavement is known. There has been so much subsequent erosion that it is not likely any striated rock pavement is exposed in Texas, though there is possibly some hope for its ultimate discovery farther south in Mexico. In

marked contrast with the exotic boulders in the Caney shale of the Ouachita Mountains in Oklahoma and Arkansas, the southern Haymond exotics are embedded in a medium apparently closely resembling the boulder clay characteristic of ice-transported deposits. The writer can recall now no essential differences between those boulders in their bed of arkosic material and the "Permo-Carboniferous" glacial deposits which he formerly studied in Brazil except that the Haymond erratics are derived from a much more diverse assemblage of rock types and the strata in which they occur are much more deformed than the nearly horizontal strata of Brazil. There can be no reasonable doubt that, if the Haymond horizon was as little deformed as the Brazilian strata, its glacial origin would be as little questioned. That there is room for legitimate doubt is readily conceded, but it is the deformation to which the Haymond and older formations were subjected that causes whatever doubt there may be.

A PETROLOGIC AND STRUCTURAL STUDY OF THE SWAKANE GNEISS, ENTIAT MOUNTAINS, WASHINGTON[1]

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ABSTRACT

The Entiat Mountains, a spur of the Cascade Range in central Washington, are composed dominantly of a metamorphic complex of pre-Ordovician age. Although originally made up of a diversified series of rocks partly of sedimentary and partly of igneous origin, metamorphism has been so complete that the original heterogeneity of the group, save that inherent in the chemical composition, has been almost completely obliterated, and the entire metamorphic mass now participates in a common foliation and forms a single definite structural unit.

Subsequent to the development of the foliation, mechanical movements superimposed cataclastic structures upon the metamorphic rocks and locally developed flat thrusts of some magnitude within them. Along the thrust surfaces mylonites are developed.

Three distinct periods of batholithic intrusion of pre-Tertiary age are recorded within the area, and these, together with the Miocene batholiths to the west, establish the presence of four periods of plutonic invasion in the Northern Cascades.

A striking fact brought out by the study is that the late Tertiary folding which trends northwest-southeast and east-west transverse to the main axis of the Cascade Range was preceded by several episodes of pre-Tertiary deformation that occurred along the same axes.

INTRODUCTION

The Entiat Mountains, located near the geographic center of the state of Washington, are a northwest-southeast–trending virgation of the Cascade Range. They are bounded on the southwest by the valley of Wenatchee River and on the northeast by the Entiat River. Both of these streams are tributary to the Columbia which crosses the Entiat Mountains antecedently about 10 miles north of the city of Wenatchee. Traced northwestward the Entiat Mountains merge into the high Cascades. Their southeastern continuation, east of Columbia River, is represented by Badger Mountain, a slightly dissected anticlinal mountain built of the Yakima basalt which underlies this portion of the Columbia Plateau.

The southern half of the Chelan quadrangle, the geology of which

[1] Presented under a different title at the meeting of the Cordilleran Section of the Geological Society of America, Pasadena, 1931.

has been recently mapped by the writer,[1] includes the eastern part of the Entiat Mountains which are underlain by an ancient gneissic complex. The study of this group of metamorphic rocks elucidates to some extent the pre-Tertiary structural history of the Cascade Range.

SUMMARY OF THE GEOLOGY OF THE SOUTHERN HALF OF THE CHELAN QUADRANGLE

The rocks of the Chelan quadrangle range in age from pre-Ordovician to Pleistocene (see map, Fig. 1). The Swakane gneiss, which forms the subject of this paper, is the oldest cartographic unit and is a complex of remarkably foliated, highly metamorphic rocks, in part of sedimentary and in part of igneous derivation.

Presumably in the late Jurassic, the area was invaded by an enormous batholith of basic granodiorite. The confines of the batholith have never been mapped, but it is known to extend over an area of more than 800 square miles. It is here called the Chelan granodiorite from the Chelan Mountains where it is typically exposed.

Following a long period of erosion the early Eocene Swauk formation,[2] composed of continental arkose, sandstone, and conglomerate made up of detritus washed from the Swakane gneiss and the Chelan batholith, was deposited. Volcanic necks of rhyolite and andesite inject the Swauk formation and the older rocks.

In the Miocene, fissure eruptions of basaltic lava built up this portion of the Columbia Plateau. In most localities the lavas overlie a hilly surface having a maximum relief of 2,000 feet, but in the vicinity of Douglas Creek they overlie or are intercalated near their base with a series of sedimentary strata bearing a fossil flora which Chaney has identified as being closely related to the Eagle Creek flora and of Lower Miocene age.[3] The lava flows advanced from the south and east, and in this district shallow lakes were formed along

[1] This paper represents a part of a thesis presented to the Graduate School of Yale University as partial fulfilment of the requirements for the Degree of Doctor of Philosophy. Many helpful suggestions of Professor Adolph Knopf, under whose direction the thesis was prepared, are gratefully acknowledged.

[2] First described by George Otis Smith from the Mount Stuart region, *U.S. Geol. Surv. Folio 106* (1904).

[3] Personal communication to Richard E. Fuller.

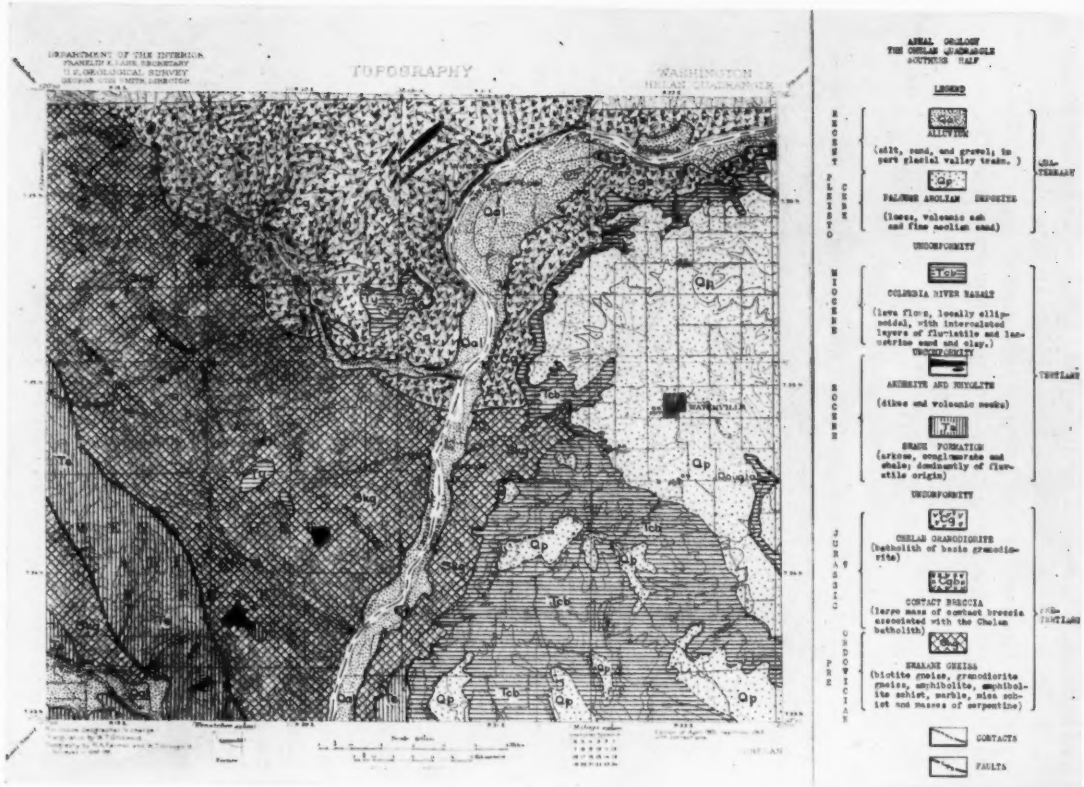

FIG. 1.—Areal geology of the southern half of the Chelan quadrangle, Washington

the margin of the lava field into which later flows advanced forming ellipsoidal lavas and palagonite tuffs.[1] Although the lavas show thin intercalations of sedimentary material throughout, near the top of the lava series there occurs a thick group of fluviatile and lacustrine sands and clays that are the equivalent of the Ellensburg formation.[2] A renewal of volcanic activity after the deposition of the Ellensburg gave rise to local masses of younger basalt. These various volcanic and sedimentary divisions have been separately mapped, but it is impossible to show them on a map of the scale of Figure 1. Accordingly they have been grouped together on this map as the "Columbia River Basalt."[3] The main thickness of basaltic lava in the Chelan quadrangle is the equivalent of the Yakima basalt and can be traced in the field into areas in the Mount Stuart quadrangle underlain by the Yakima lavas.

In the Pleistocene the plateau area east of Columbia River was mantled with a deep covering of aeolian deposits, principally loess and volcanic ash.

DEFINITION, OCCURRENCE, AND AGE OF THE SWAKANE GNEISS

The metamorphic rocks of the Entiat Mountains occur in two northwest-southeast–trending belts separated by a graben floored with highly deformed strata belonging to the Swauk formation (Fig. 1). The rocks in the smaller belt are overlain to the southward by the Swauk strata of the Wenatchee Valley. The larger northern belt is terminated in the map area by the intrusive contact of the Chelan batholith, this contact following roughly the course of Mad

[1] Richard E. Fuller, "The Aqueous Chilling of Basaltic Lava on the Columbia River Plateau," *Amer. Jour. Sci.* (5), Vol. XXI (1931), pp. 281–301.

[2] *U.S. Geol. Surv. Folio 86* (1903).

[3] Under the name "Columbia River lava," I. C. Russell included all basaltic flows of the Columbia River Plateau ranging in age from Eocene to Pleistocene ("Geology and Water Resources of Nez Perce County, Idaho," *U.S. Geol. Surv., Water Supply Paper 53* [1901], p. 28). Later Merriam restricted the term "Columbia lava" to include only flows of Miocene age ("A Contribution to the Geology of the John Day Basin," *Univ. Calif. Publ., Bull. Dept. Geol. Sci.*, Vol. II [1901], p. 303). The usage of Russell has been more generally followed and seems more desirable particularly since the Miocene lavas are called the Yakima basalt in the numerous U.S. Geol. Surv. reports dealing with the Columbia Plateau in Washington (George Otis Smith, "The Ellensburg Folio," *U.S. Geol. Surv. Folio 86* [1903]).

FIG. 1.—Areal geology of the southern half of the Chelan quadrangle, Washington

River and the lower Entiat River. The metamorphic rocks formerly had a much greater extension to the north, however, and are frequently found as roof pendants in the Chelan batholith. The extension of the metamorphic rocks into the high Cascades to the northwest has not been mapped, or even traced in a reconnaissance way. Their eastern continuation is concealed by the lavas of the Columbia Plateau.

In the Chelan quadrangle the finest exposures of this group of metamorphic rocks occur along the canyon of Swakane Creek from which they are accordingly named. The foliation of the rock at this locality is almost horizontal (Fig. 2), and the precipitous canyon walls afford natural cross-sections rising 3,000 feet above the stream on either side.

FIG. 2.—Horizontally foliated biotite gneiss. Lincoln Rock, mouth of Swakane Canyon. (Photo copyrighted by A. G. Simmer.)

The Swakane gneiss is overlain nonconformably by the Swauk formation of early Eocene age and it is intruded by the Chelan batholith. A few miles southwest of the area there are outcrops of a group of slates with interbedded masses of chert and limestone, the Peshastin formation of George Otis Smith.[1] Warren S. Smith[2] correlates the Peshastin formation with a series of sedimentary rocks containing definite Trenton fossils that occur in the Skykomish basin to the west. On the basis of comparative metamorphism the Swakane gneiss is therefore of pre-Ordovician age.

PETROGRAPHIC CHARACTER OF ROCK VARIETIES

The metamorphic rocks vary greatly in petrologic character. Although a fine-grained, well-foliated biotite gneiss predominates, many other rock types are included. Coarse mica schist, crystalline marble, amphibolite, amphibolite schist, various lime-silicate rocks, and rudely laminated granodiorite and quartz diorite gneisses are found

[1] G. O. Smith, "Mount Stuart Folio," *U.S. Geol. Surv. Folio 106* (1904).

[2] W. S. Smith, "Stratigraphy of the Skykomish Basin, Washington," *Jour. Geol.*, Vol. XXIV (1916), pp. 559–83.

and locally form areas of considerable size (see Fig. 3). Despite this varied assemblage of different rock types, the observer is impressed in the field more by the homogeneity of the group than by its heterogeneity. All of the varieties participate in a common foliation, and the majority of them are transitional into one another, rendering accurate mapping into separate definite units impossible. It is clear that they have been produced by metamorphic agencies operating regionally on a series of rocks of diverse chemical composition and geologic origin. Metamorphism has been so complete that the original structural diversity has been largely obliterated.

METAMORPHIC ROCKS OF SEDIMENTARY ORIGIN

The staple rock of the Swakane gneiss is a gneissic rock of uncertain origin composed of feldspar, quartz, and both micas. Locally intercalated with these gneisses are small masses of metamorphic rock which are definitely of sedimentary derivation. The most widespread is a series of marble lenses and associated lime-silicate rocks which occur for the most part in the northwestern corner of the area (Fig. 3) near the border of the Chelan granodiorite batholith. At nearly all localities the lenses are concordant with the lamination in the surrounding gneiss, but locally their trend deviates at a small angle. Associated with the marble are other metamorphic rocks of sedimentary origin, the most conspicuous of which is a mica schist containing numerous small quartz lenticles.

Two varieties of marble are present. The more common is a pure white coarsely crystalline variety which the microscope shows to be almost pure calcite with a few grains of plagioclase (An. 38) and a small amount of clinozoisite. Near the contact with the batholith this rock locally becomes rich in tremolite and epidote and may be traversed with numerous quartz-epidote veinlets.

At other localities the marble is a striking and quite unusual rock due to the fact that it is of such fine grain that it has an aphanatic, and sometimes almost a glassy, appearance. The microscope reveals little more than a fine mat of calcite granules with an average grain size of 0.001 mm. Light green laminations occur locally in which andesine (An. 44), clinozoisite, and minute grains of an obscure, feebly birefracting greenish-yellow mineral are fairly common.

Fine-grained marble is found as a small intercalation within the

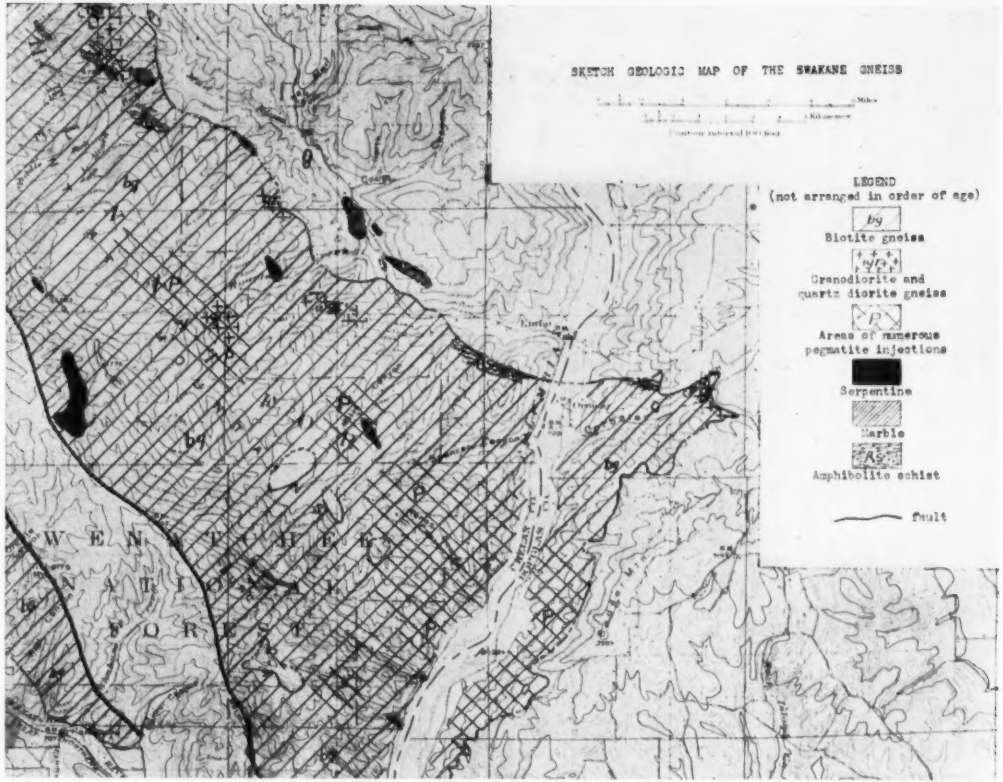

FIG. 3.—Geologic map of the Swakane gneiss

gneiss about 1½ miles south of the mouth of Swakane Creek. An analysis of this marble, given in *Washington Geological Survey Bulletin 4*, is reprinted below:

SiO_2	0.38
Al_2O_3 and Fe_2O_3	0.21
CaO	55.26
MgO	none
Loss on ignition	43.98

The analysis confirms the microscopic evidence that the rock is nearly pure calcium carbonate.

Gneiss derived from impure calcareous sediments.—Associated with the layer of marble south of the mouth of Swakane Creek occur belts of gneiss which show by their mineralogical composition that they were derived from sediments containing appreciable amounts of silica and alumina as well as lime. Biotite, garnet, quartz, and feldspar are megascopically visible, and the rock reacts to acid showing the presence of carbonates.

Under the microscope the rock is of gneissose structure and contains a wide variety of constituent minerals. Quartz and plagioclase (An. 38) are the most abundant minerals and form granular bands separated by schistose lenticles of biotite, hornblende, and calcite. Garnet and euhedral crystals of clinozoisite are also abundant. Other minerals, named in decreasing order of abundance, are orthoclase, titanite, magnetite, and zircon.

Rocks of this same general type can be found locally adjacent to the previously described belt of marble in the northwestern part of the area. A highly quartzose muscovite schist is also associated with this belt near the headwaters of Hornet Creek. In the field at both localities a complete and very gradual transition can be found between the lime-silicate rocks and the biotite gneisses that make up the bulk of the Swakane gneiss.

FIG. 3.—Geologic map of the Swakane gneiss

METAMORPHIC ROCKS OF IGNEOUS DERIVATION

Gneisses definitely derived from plutonic masses of granodiorite and diorite, as well as amphibolite and amphibolite schist derived from basic igneous rocks, are found at several localities. Excluding the amphibolite schist they make up only an insignificant fraction

of the total area of metamorphic rocks, though exceeding slightly the metamorphic rocks definitely of sedimentary origin. Gneisses derived from intermediate and acidic plutonic rocks are found in the upper portion of Corbaley Canyon, along Roaring Creek, and at several other localities whose positions are plotted on the map (Fig. 3). Amphibolite is of rare occurrence and is confined to very small masses. Amphibolite schist is very abundant along the border of the belt near the contact with the Chelan batholith and is commonly encountered farther north as roof pendants in the batholith.

Granodiorite gneiss and quartz diorite gneiss.—Just above the forks of Corbaley Creek the canyon walls are composed of a well-foliated to obscurely banded mass of metamorphosed plutonic rock ranging in composition from basic granodiorite to quartz diorite. The foliation, though everywhere clear in the field, decreases up the canyon, and this change is accompanied by an increase in the grain size and in the quartz content of the rock. Immediately below the contact with the Yakima basalt, the rock is of coarse-grained granitic appearance with only a faint foliation. White feldspar, biotite, hornblende, and quartz are recognizable megascopically. Aside from the obscure foliation and some visible evidence of cataclastic deformation, hand specimens of the rock closely resemble those from the Chelan granodiorite batholith. In thin section the rock is found to be composed of basic oligoclase (An. 30), quartz, biotite, hornblende, orthoclase, titanite, apatite, and magnetite, named in order of decreasing abundance. The texture bears a much closer resemblance to the hypidiomorphic granular texture of intrusive rocks than to a metamorphic texture, although the rock does show evidence of cataclastic deformation and of a limited amount of recrystallization. The plagioclase of the rock is never zonal, and this fact, together with the absence of augite and the presence of cataclastic effects, easily differentiaties the rock from the Chelan granodiorite.

The dioritic varieties, exposed lower in the canyon, are more distinctly foliated, and thin sections show that they have suffered far more recrystallization. Some thin sections show a curious interlamination of crystalloblastic and hypidiomorphic granular bands, suggesting that recrystallization followed some previous directional structure in the rock. Comparison with the less recrystallized grano-

diorite gneiss suggests that this earlier structure may have been produced by cataclastic deformation. In some of the other granodiorite gneisses, notably the mass on Roaring Creek, it can be shown that cataclastic effects were produced in the rock at the same time or prior to recrystallization. This early cataclastic deformation, visible only in the granodiorite and quartz diorite gneisses, is entirely separate and distinct from the widespread cataclasis to be subsequently described as affecting all rock units in the Swakane gneiss.

In passing down Corbaley Canyon seemingly every transition is found between these plutonic rocks and the biotite gneisses that make up the main volume of the Swakane gneiss. The biotite gneiss at the foot of the canyon is considerably richer in quartz than the plutonic rocks at the head of the canyon, however.

The gneiss outcropping near the headwaters of the first canyon south of Roaring Creek is of similar nature. It shows marked recrystallization effects under the microscope, and the outcrops in the field are well foliated.

The gneiss derived from acidic plutonic rock outcropping on the South Fork of Roaring Creek is a very well-foliated granodiorite gneiss containing a small amount of muscovite. In hand specimen the rock has a knotty appearance caused by the warping of biotite flakes around lenticular masses of feldspar representing original crystals of the plutonic rock which were crushed but not recrystallized. Despite the presence of these numerous cataclastically deformed porphyroclasts, the texture of the rock is controlled largely by recrystallization. Two episodes of cataclasis are seen: the older one affecting only the porphyroclasts of feldspar which are surrounded by recrystallized matrix, and the more recent, which also affects all other rock varieties in the Swakane gneiss, being seen both in groundmass and porphyroclasts.

The contacts of these metamorphosed igneous masses with the normal biotite gneiss that makes up the bulk of the Swakane group are always transitional, but locally the transition occurs more abruptly than at the Corbaley Canyon locality.

Amphibolite schist.—Along the northeastern border of the belt of metamorphic rocks, amphibolite schist and locally coarse-grained

amphibolite gneiss are present at several localities (see map, Fig. 3).

Wherever exposed, the rock is invariably well foliated. Megascopically, small well-oriented crystals of hornblende are the most abundant constituent. Feldspar is also recognizable in all specimens. Bands made up of granular feldspar and epidote are interlaminated with normal schist at some localities.

A thin section from a specimen collected in the short canyon southwest of Entiat shows the rock to have a typically open schistose texture made up of well-oriented hornblende and titanite surrounding roughly equidimensional crystals of feldspar, epidote, and rarely quartz. The mineral constituents named in order of decreasing abundance are hornblende, andesine (An. 34), epidote, quartz, titanite, orthoclase, magnetite, apatite. Hornblende and feldspar make up more than 90 per cent of the rock.

A specimen taken from a large roof pendant southeast of the mouth of Roaring Creek is of similar nature, but contains a considerable amount of quartz. A slide cut from a specimen collected east of the mouth of Indian Creek is unusual for its high content of titanite which forms 4 or 5 per cent of the total mineral composition. Titanite is an abundant accessory in the amphibolite schist from all localities but rarely exceeds more than 1 per cent of the total volume.

The areas of amphibolite schist probably represent metamorphosed basic lava flows which were intercalated in the sedimentary series that furnished the marble, mica schists, and other metamorphic rocks of sedimentary origin. Their mineralogical composition is compatible with that of a recrystallized basalt, but metamorphism has been so thorough that all vestiges of the original volcanic structure, if once present, have been destroyed.

METAMORPHIC ROCKS OF UNCERTAIN ORIGIN

The rocks comprising the main bulk of the Swakane gneiss have been so thoroughly transformed that their original derivation is in doubt. The relations are further complicated by the fact that in the field certain gneisses of distinctive character appear to be transitional directly into both metamorphic rocks of igneous origin and into metamorphic rocks of sedimentary origin.

Biotite gneiss.—Over 85 per cent of the area occupied by the Swakane gneiss is composed of a biotite gneiss of rather uniform composition and structure. The rock is exceedingly well foliated and of fine grain. Indeed, some of the finer-grained varieties may be classified either as coarse schist or fine-grained gneiss. Megascopically, biotite is one of the most prominent minerals, but its conspicuousness is due largely to the fact that its flat flakes cover a large percentage of the cleaved surfaces of the rock. The biotite shows the deep reddish-brown pleochroism regarded by Becke as characteristic of the mica in metamorphic rocks from the deepest zone. Quartz and feldspar are always present, and colorless mica and garnet can be found in varying quantities in practically all specimens. At some localities muscovite exceeds biotite in amount. In thin section quartz and feldspar are found to be present in about equal proportions. Both feldspars are present, but a plagioclase (commonly a calcic oligoclase) usually predominates. In addition to muscovite and garnet, titanite, zircon, apatite, and magnetite are usually present in small amounts.

In all specimens of the biotite gneiss the texture of the rock has been controlled wholly by recrystallization. The minerals are fresh and clear, and they show the allotriomorphic and mutually interpenetrating boundaries of a typical crystalloblastic rock.

A representative specimen of the biotite gneiss from the east wall of the Columbia gorge opposite the mouth of Swakane Creek was measured by the Rosiwal method with the following results:

Mineral	Volume Percentage
Quartz	43.9
Feldspar	41.8
Biotite	13.3
Muscovite	0.7
Accessories	0.3
Computed specific gravity	2.7
Determined specific gravity	2.67

The feldspar in this rock is largely plagioclase (An. 29). The specimen shows a greater preponderance of biotite over muscovite than is commonly the case.

With regard to the original nature of the biotite gneiss, the high percentage of quartz suggests that it is a transformed sediment, and this suggestion is in harmony with the occurrence of layers of marble and of lime-silicate rocks interlaminated with it. On the other hand, masses of granodiorite gneiss can be found which appear to be transitional with the biotite gneiss, and the high feldspar content of the biotite gneiss is difficult to explain on the assumption that it is a transformed sediment. The biotite gneiss shows local variation in composition, particularly with regard to the relative amounts of biotite and muscovite and of feldspar and quartz. Such aberrant rock types as are produced by these variations almost invariably occur in definite belts that can be followed along their trend for considerable distances. This feature is most readily explainable as being due to original diversity in sedimentary beds. A sedimentary derivation is also favored by the fact that the marble and other rocks definitely classifiable as of sedimentary origin commonly form interbeds of considerable linear extent but of small thickness within the gneiss. Because of these features it is considered most probable that the biotite gneiss represents a series of metamorphosed sediments probably largely of arkosic composition.

If we accept this view of the origin of the biotite gneiss, the remarkable transition between it and the granodiorite gneiss may have occurred through the deposition of arkose derived from the granodiorite directly on the eroded surface of the plutonic mass followed by dynamic metamorphism of both rocks. In this way the peculiar transition, with the rock increasing in quartz as the normal biotite gneiss is approached, could have been produced.

The contact between biotite gneiss and granodiorite gneiss is not interwoven, and it is therefore unlikely that the granodiorite gneiss is the younger and that the transition represents an injection zone.

It might be suggested that the granodiorite gneiss is a palingenetic eruptive[1] formed by ultrametamorphism of the biotite gneiss. Under this hypothesis, however, the granodiorite gneiss should be higher in the alkalies and in silica than the parent biotite gneiss,

[1] J. J. Sederholm, "On Migmatites and Associated Pre-Cambrian Rocks of Southwestern Finland," Part 1, "The Pellinge Region," *Bull. Comm. Geol. Finlande No. 58* (1923), and Part 2, "The Region around the Barosundsfjard West of Helsingfors and Neighboring Areas," *ibid., No. 77* (1926).

whereas exactly the reverse relationship actually exists. In addition, the contacts should be interwoven like those of an injection zone instead of being gradually transitional.

INTRUSIVE ROCKS IN THE SWAKANE GNEISS OLDER THAN THE CHELAN BATHOLITH

In some portions of the area occupied by the Swakane gneiss the metamorphic rocks are literally riddled with subsequent igneous injections of small size. Aside from several stocklike masses of ultrabasic rock now altered to serpentine, the majority of these injections are of pegmatitic composition, but lamprophyres and related dikes are also present. Veins of milky quartz practically devoid of metallic constituents are also associated with the areas that contain numerous pegmatites and are undoubtedly related to the same period of igneous activity.

Pegmatite.—Although injections of pegmatite are distributed throughout the area occupied by the Swakane gneiss, they are comparatively rare in the belt of gneiss that lies south of the graben, and in the area adjacent to the contact of the Chelan batholith. Along Columbia River in the central portion of the gneissic belt they are present in enormous numbers. On the east bank of the Columbia where it transects the Entiat-Badger Mountain divide and in the lower part of Tenas George Canyon, pegmatite makes up from 15 per cent to 25 per cent of the rock exposed. Pegmatite injections are also very common along the north and south forks of Roaring Creek and in the upper part of Mills Canyon. An area containing great numbers of quartz veins is found on the eastern slope of Burch Mountain, but pegmatites are comparatively rare at this locality. The areas in which pegmatites are particularly abundant are shown by special convention on the sketch map of the Swakane gneiss (Fig. 3).

In field habit the pegmatites are small, sill-like injections along the planes of lamination in the gneiss. The sills are never of great size, 6 or 10 inches being an average figure for their thickness. Typical exposures on the east bank of the Columbia opposite the mouth of Tenas George Canyon show a very uniform alternation of gneissic layers 1 to 5 feet in thickness, with pegmatites having an average

thickness of 6 to 8 inches. The planes of gneissic banding at this locality, and also at all other localities where pegmatite intrusions are particularly abundant, lie practically horizontal, and consequently the high cliffs of the mixed rock have a very striking appearance due to the alternation of thin bands of almost pure white pegmatite and relatively dark-colored gneiss. The pegmatite sills are very straight and do not vary appreciably along their strike so that such cliffs have a ruled or ribboned effect. In this region these cliffs are popularly referred to as the "Ribbon Rocks."

The continuity of some of the pegmatite injections is remarkable. Individual sills only 6 inches in thickness can be traced for more than 200 yards without showing appreciable variation in size.

The contacts between the pegmatites and the adjacent gneiss are invariably sharp and clean cut. There is absolutely no trace of interaction between the pegmatite injections and the gneiss such as occurs in migmatites, which the rocks otherwise resemble.

Megascopically the most abundant constituent of the pegmatites is a white feldspar. In many injections it is intergrown with quartz to form graphic granite. Biotite and muscovite are generally present in large hexagonal plates, the dark mica commonly predominating. Garnet is a local accessory. Under the microscope the feldspar proves as a rule to be microperthite, although in some specimens unintergrown orthoclase is present. Practically all specimens are cataclastically deformed.

In a few examples a variation in texture between the central and marginal portion of a pegmatite sill was noted. The margins are of aplitic nature, while the central part is of coarse pegmatitic structure and is considerably enriched in quartz over the marginal portion. This characteristic of the pegmatites in the Columbia Valley was also noted by Smith and Calkins, who found one example in which the central filling was of quartz.[1] Multiple and composite injections of pegmatite are common, but, in the cases previously noted, the variation in composition is due to local differentiation and not to separate intrusion.

Certain peculiarities of the distribution of the pegmatite-rich areas must be taken into account in any explanation of their genesis:

[1] *U.S. Geol. Surv. Bull. 235* (1904), p. 76.

1. The pegmatite-rich areas lie in a northwest-southeast–trending linear zone, occupying the central portion of the larger belt of the Swakane gneiss.

2. The pegmatites occur in great numbers only where the planes of lamination in the gneiss are horizontal or have low dips.

3. They are abundant only where dissection has cut deeply into the gneiss. Pegmatites are not present in great quantity on the ridge between Roaring Creek and Mills Canyon, or in other parts of the Entiat Range at high elevation.

4. They show no preference for, or relation to, the regions occupied by either the granodiorite gneiss facies of the Swakane gneiss or by the Chelan batholith.

This last conclusion, if tenable, is important because it establishes the presence of three epochs of batholithic intrusion in the Chelan quadrangle, two of which, being older than the Chelan batholith, are probably pre-Jurassic. Several lines of evidence indicate that the pegmatites definitely belong to a younger epoch of igneous activity than the granodiorite gneisses included in the Swakane group:

1. The granodiorite gneisses are well foliated in most localities due to recrystallization under pressure subsequent to their consolidation. The pegmatites are commonly cataclastically deformed, but show no evidence of recrystallization. On the South Fork of Roaring Creek, undeformed pegmatites intrude in *lit-par-lit* manner a granodiorite gneiss that is well foliated and almost completely recrystallized. The pegmatites were intruded after the foliation had been produced and show no evidence of recrystallization (Fig. 4).

2. The pegmatites show no tendency to increase in number toward the granodiorite gneiss or quartz diorite gneiss masses. On the other hand, the pegmatite-rich areas and the areas of plutonic gneiss occur typically in different parts of the area.

3. The pegmatites definitely intrude all facies of the Swakane gneiss with which they have been observed in contact. The granodiorite gneisses appear to be transitional with the other facies of the Swakane group.

The evidence that the pegmatites are older than the Chelan batholith is even more conclusive. The lack of any spatial relation of the pegmatites to the batholith may be cited as evidence against a genet-

ic connection between them. The pegmatites show pronounced cataclastic structures which are not seen in the Chelan granodiorite. Also at a locality north of Corbaley Canyon and again on the ridge south of Roaring Creek the Chelan batholith cuts across pegmatite stringers which in turn intrude the gneiss.

FIG. 4.—*Lit-par-lit* injections of pegmatite cutting foliated granodiorite gneiss. South Fork of Roaring Creek.

Because of the relations outlined above, the pegmatites are believed to be related to a large subjacent igneous body which erosion has not yet fully revealed. This hypothesis accounts for the fact that extensive areas of pegmatite occur only where erosion has cut

deep incisions into the Swakane gneiss as at the localities in the Columbia River Canyon, in the upper part of Mills Canyon, and along Roaring Creek. The parent igneous body, probably a granite batholith, was evidently intruded after the foliation in the Swakane gneiss was produced, but prior to the cataclastic deformation.

Serpentine.—The position of four small stocklike masses of serpentine enclosed in the Swakane gneiss are indicated in Figure 3. Other masses of more modest dimensions were also encountered.

Microscopic study of the better-preserved masses shows that the serpentine was derived from intrusions of saxonite, although some of the stocks have been so thoroughly altered that the original mineral composition of the intrusive rock cannot be definitely ascertained. At the present time, talc and bastite are the principal minerals, although magnetite and other accessories are locally common. Veinlets of asbestos form an anastomosing network throughout the rock. Although the central portions of the bodies of serpentine are commonly structureless, a faint foliation is found along their margins, and in this foliated portion talc is very abundant.

Locally, near the border of the Chelan batholith the Swakane gneiss contains small bodies of rock made up dominantly of talc and serpentine minerals but also containing appreciable amounts of tremolite and occasionally epidote. These masses are probably related to the serpentine, but their mineralogical composition and textures do not preclude the possibility that they represent metamorphosed impure masses of dolomite or similar magnesia-rich rock. Whatever their original nature, their metamorphism was certainly not synchronous with that of the Swakane gneiss, for their principal mineral, talc, is a product of low temperatures.

ATTITUDE AND ORIGIN OF THE GNEISSIC FOLIATION

In the well-exposed natural cross-section made by the Columbia River Canyon it is readily seen that the attitude of the gneissic foliation conforms to a broad, flat-topped anticline. The axis of this anticline corresponds roughly with the crest of the present Entiat Mountains. Locally, the foliation for short distances does not conform to the anticlinal structure, but all deviations observed may be logically related to local deformation adjacent to the intrusive con-

tact of the Chelan batholith or adjacent to the faults that define the graben in the southwestern part of the area. The foliation on the limbs of the anticline may locally dip at angles of 70°, but the area of horizontal foliation along the axis is very broad.

The origin of the gneissic foliation, and particularly its development in an anticlinal structure, is difficult to explain. The exceptional development of the foliation and the occurrence of interlaminated bodies of marble and mica schist, following the planes of gneissic banding, suggest that the foliation may have been produced through replacement of the stratification of the original sedimentary rocks by the growth of new minerals oriented along the bedding planes: a process which Sander calls "*abbildungskristallization.*" Certain of the gneisses derived from calcareous sediments that lie adjacent to the bands of marble in the Swakane gneiss are very similar to the marbles of Macedonia which Erdmannsdorffer[1] believes owe their origin to this process. If the foliation were developed in a horizontal position through this process the subsequent uparching of the metamorphosed rocks into an anticlinal structure might account in part for the cataclastic deformation which is so conspicuous in all rock units of the Swakane gneiss. There are, however, several difficulties encountered in explaining the foliation of the gneiss as a replacement structure. The trend of the belts of marble does not always correspond to the strike of the foliation, but in places deviates from it at a small angle. This discordance in trend is well shown by the belt of marble outcropping on Gold Ridge. Also the uniform passage of the foliation across rocks such as the granodiorite gneiss which supposedly did not have a directional structure prior to metamorphism would indicate that the gneissic banding was produced by a different process. The granodiorite gneiss outcropping along the South Fork of Roaring Creek, and the area of granodiorite gneiss lying east of the same creek, is as well foliated as the normal biotite gneiss. Possibly an earlier directional structure was produced in these formerly igneous masses by cataclasis and then largely obliterated by subsequent recrystallization. The presence of palimpsest cataclastic structures in some of the granodiorite gneisses has already been commented upon.

[1] O. H. Erdmannsdorffer, "Untersuchungen an mazedonischen Gesteinen," *Neues Jahrb., Beilage-Band 48* (1923), pp. 75–113.

In many pre-Cambrian intrusive masses,[1] and in some of later date,[2] a marked foliation has developed around the contacts of the intrusive mass. In the Swakane gneiss the areas of numerous pegmatite injections and the axis of the anticlinal structure roughly coincide. This suggests that the archlike structure in the gneiss may be due to the development of peripheral foliation around a subjacent igneous body of batholithic or laccolithic shape which erosion has not yet fully revealed and of which the pegmatites are the principal surface manifestation. Concordant batholiths of this type generally contain many inclusions and are bordered by extensive injection zones. There are, however, grave difficulties encountered in explaining the pegmatite-rich areas in the Chelan quadrangle as an injection zone of this nature. As has been previously pointed out, the contacts between pegmatite and gneiss are always absolutely sharp and clean cut. There is no evidence whatever of reaction between the pegmatitic magma and the adjacent wall rock to form migmatites. If the intrusion of the pegmatites was synchronous with the recrystallization that produced the gneiss, some interaction between the magma and the adjacent wall rock would be expected. At a point near the confluence of the Chelan and Columbia rivers where the Chelan batholith is intrusive into amphibolite schist belonging to the Swakane group of rocks, true migmatites have been produced, and these rocks are of strikingly different character from the pegmatite-riddled gneisses in question.

If the foliation of the gneiss were developed in a horizontal position, its subsequent uparching into an anticlinal structure may be attributed in large part to orogenic uparching of the Entiat Mountains in Tertiary and pre-Tertiary time. The Columbia River basalt arches over Badger Mountain in a broad anticline whose axis roughly corresponds to the anticlinal axis in the gneissic foliation. Also the early Eocene Swauk formation which floors the Wenatchee Valley is tilted away from the Entiat Mountains at angles of 55°–90°, thus testifying to the fact that the mountains were an axis of uplift subsequent to the deposition of the Swauk, but prior to the outpouring of the Yakima basalt.

[1] F. D. Adams and A. E. Barlow, "Geology of the Haliburton and Bancroft Areas," *Can. Geol. Surv. Mem. 6* (1910).

[2] H. W. Turner, "Bidwell Bar Folio," *U.S. Geol. Surv. Folio 43* (1898).

On the other hand, if the gneissic foliation dipped steeply at the time of its development, as is more commonly the case with dislocation metamorphic rocks, it is much more difficult to explain the present anticlinal structure by subsequent warping and fracturing. Although the pronounced cataclasis which affects all units of the gneiss testifies to considerable post-foliation movement, it seems hardly possible that the anticlinal structure could have been produced from a mass of steeply dipping gneiss by later mechanical movements.

To the writer both the origin of the foliation and its present anticlinal structure remain a complete enigma.

CATACLASTIC DEFORMATION

After the development of the foliation in the Swakane gneiss, marked mechanical movements superimposed cataclastic structures on the older foliation. This cataclastic structure is perceptible in thin sections taken from almost any point in the gneissic area, but the amount and character of the cataclasis varies. The most extreme results are found along the western wall of the Columbia gorge adjacent to Burch Mountain where the biotite gneiss along the surfaces of a series of flat thrusts has been ground and powdered into mylonite. In most localities, however, little evidence of the deformation is present in the field although cataclastic effects may be marked in thin section. In a specimen taken from the extreme headwaters of the first creek south of Roaring Creek, the only field indication of cataclasis is a slightly more sugary aspect of the surface of the rock than is exhibited in the usual outcrops of biotite gneiss. In thin section, however, it may be seen that the mineral grains are much granulated and broken, especially along their edges (Fig. 5). All of the larger quartz grains show strain shadows. Microscopic shearing and displacement of the albite lamellae in the plagioclase is very conspicuous under crossed nicols. The edges of the biotite grains are ragged and have been irregularly penetrated by fragments broken from the other minerals, and the folia of biotite are badly bent and twisted. The cataclasis is clearly subsequent to the foliation. The deformation has been entirely mechanical; there is no trace of the formation of new minerals.

The cataclasis affects all rock units of the Swakane gneiss as well

as the pegmatites which intrude it. Cataclastic effects have not been observed in the Chelan granodiorite batholith, however, nor are they perceptible in the stocks of serpentine unless the production of talc in these masses represents the equivalent response to the deforming forces.

The mechanical movements that gave rise to the cataclastic structure also broke the gneiss and caused horizontal displacements

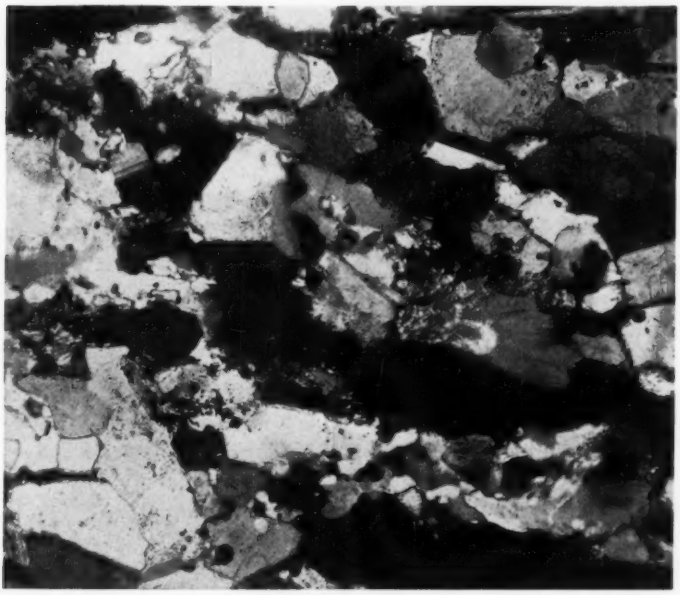

FIG. 5.—Photomicrograph of cataclastically deformed biotite gneiss from Roaring Creek. Note the granulation along the mineral borders. (Ordinary light, ×60.)

within it along a series of flat thrusts. The best developed of these thrusts is well exposed along the Columbia on the eastern slope of Burch Mountain. Although one nearly flat surface appears to have accommodated a large part of the movement, several parallel fault surfaces separated by thin plates of gneiss, as well as a few steeply dipping faults, are associated with it. South of Swakane Creek, however, the main fault is rendered especially conspicuous by a

discordance in dip of the gneissic foliation above and below the fault surface. This general zone of faulting can be traced continuously from a point near the southern border of the area to the mouth of Tenas George Canyon. It extends farther in both directions, but is lost at each locality beneath a covering of rocky soil. The discordance in dip between the lamination in the blocks on the two sides of the thrust decreases until near Tenas George Canyon the lamination in the gneiss both above and below the thrust surface is essentially horizontal. The amount of displacement along the thrust is not known, since only the Swakane gneiss is involved in the movement, and definite stratigraphic zones are not available in the gneiss by which to measure the magnitude of the movement.

Along the thrust surface south of the mouth of Swakane Creek there is a zone of rock varying from a few inches to 10 or 12 feet in thickness which contrasts strikingly with the gneiss above and below the thrust surface. This rock is of a dirty yellowish-brown color and shows a pronounced lamination at most localities. In the field it closely resembles a badly altered flow of rhyolite. Microscopically, however, it is seen to be a well-laminated microbreccia, or mylonite, which has been produced by the comminution and pulverization of the gneiss along the thrust plane. Every transition between this mylonite and the gneiss from which it was derived can be seen in the field without the aid of the microscope. Locally the mylonite contains eyelike inclusions of the gneiss which have been badly crushed and granulated, but not completely reduced to powder. In some places the microbreccia passes into coarser breccias made up of gneissic fragments averaging a centimeter or less in diameter, but which clearly show their gneissic origin even in hand specimen. The lamination of the mylonite reminds one forcibly of the flow structure of acidic igneous rocks. The lamination is, in general, parallel with the thrust surface but it is wavy and irregular in detail. The flow bands bend around fragments that have not been completely crushed and also become very irregular and contorted at localities where the mylonite zone enlarges or thins rapidly.

Interesting effects are seen in the rocks adjacent to the mylonite zone. Pegmatites are locally common in the area cut by the thrust, and practically every one of them has been reduced to a granular

breccia or kakirite,[1] even though the gneiss in which the pegmatites lie has not been appreciably brecciated. Once reduced to this brecciated condition, however, the pegmatites resist further mylonitization. Another interesting phenomenon is the development of cleavage in the quartz veins adjacent to the thrust zone. Several quartz veins, near the point where they are cut by the thrust, can be split

FIG. 6.—Photomicrograph of mylonite from Burch Mountain. Note the feldspar crystal broken into thousands of pieces, at the extreme right of the slide. (Ordinary light, ×60.)

into plates a fraction of an inch in thickness and several inches in length. The cleavage is not apparent some distance from the thrust surface.

Microscopically the mylonite is remarkable for its complete absence of structures due to recrystallization. The deformation of the rock has been entirely mechanical, and practically no new minerals

[1] Percy Quensel, "Zur Kenntnis der Mylonitebildung, erläutert an Material aus dem Kebnekaisgebiet," *Bull. Geol. Inst. Upsala*, Vol. XV (1916), pp. 91–117.

have formed. In this respect the mylonites are very different from those associated with the Caledonian thrusts in the Northwest Highlands of Scotland and in Scandinavia. They are, in some respects, similar to the mylonites in Alsace described by Jung,[1] but even these rocks contain small amounts of quartz, feldspar, and other minerals of new formation. In the mylonites from the thrust

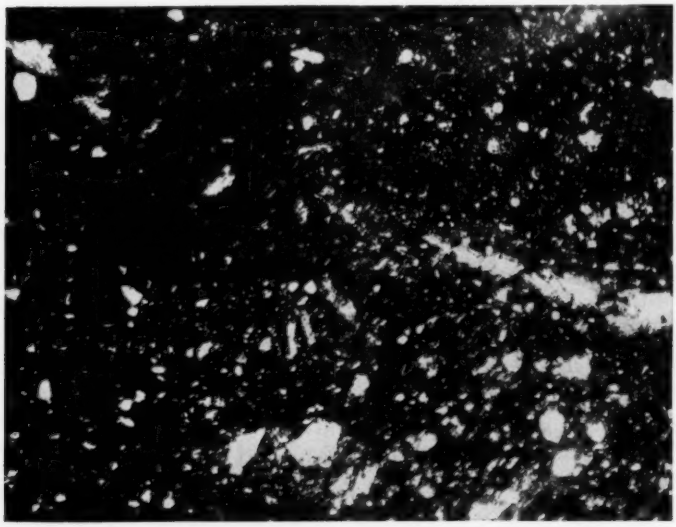

FIG. 7.—The same view as Figure 6 under crossed nicols

on Burch Mountain the only new minerals are minor amounts of calcite and chlorite.

In thin sections taken from various parts of the mylonite zone, every stage is visible in the demolishing of the gneiss into a fine powder scarcely resolvable under the microscope. In the finer grained rocks thin sections show no porphyroclasts, and the rock consists of a dark-colored base made up of minute particles averaging less than 0.05 mm. in diameter set with numerous larger particles ranging up to 0.3 mm. These larger particles are generally feldspar,

[1] J. Jung, "Contribution à la geologie des Vosges Hercyniennes d'Alsace," *Mem. du service de la carte geologique d'Alsace et de Lorraine, No. 2* (1928), pp. 158–71.

less commonly quartz. The rock grades into varieties of the same general appearance which contain, however, streaky porphyroclasts of badly crushed gneiss. The rocks containing occasional porphyroclasts are the most common varieties in the mylonite zone. The lamination, which is so characteristic of the rocks in the field, is also visible in thin section, but in most specimens is not very pro-

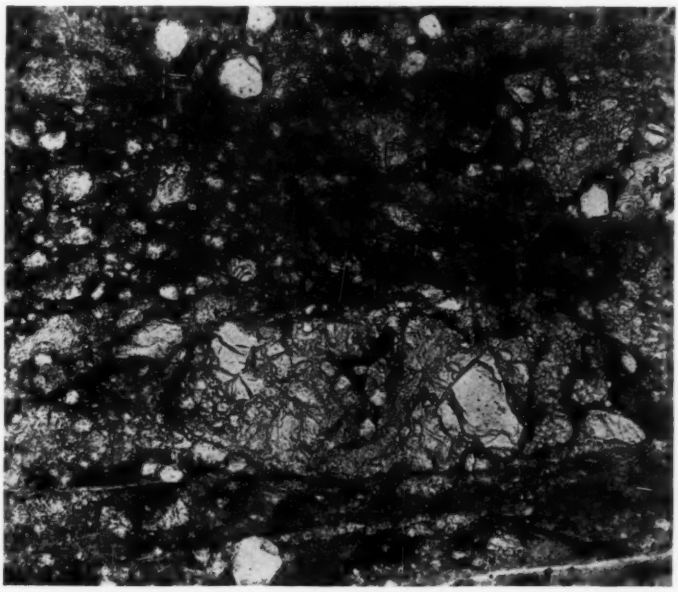

FIG. 8.—Photomicrograph of a badly crushed porphyroclast in mylonite from Burch Mountain. Note flowage phenomena in the powdery groundmass and the rounded character of the larger quartz grains. (Ordinary light, ×60.)

nounced. In some rocks of this zone especially badly crushed porphyroclasts show lamination which does not always lie in the same plane as the lamination of the base of the rock, indicating that the porphyroclasts have been rolled and revolved. With increase in number of porphyroclasts the mylonites gradually merge into breccias or kakirites whose nature is megascopically visible. Adjacent to some individual porphyroclasts every transition can be seen from

the original gneiss into the laminated rocks of ultramylonitic character first described.

Small mylonite zones occur associated with thrusts in other parts of the area, but the microbrecciation is not as well developed as at the locality on the east slope of Burch Mountain.

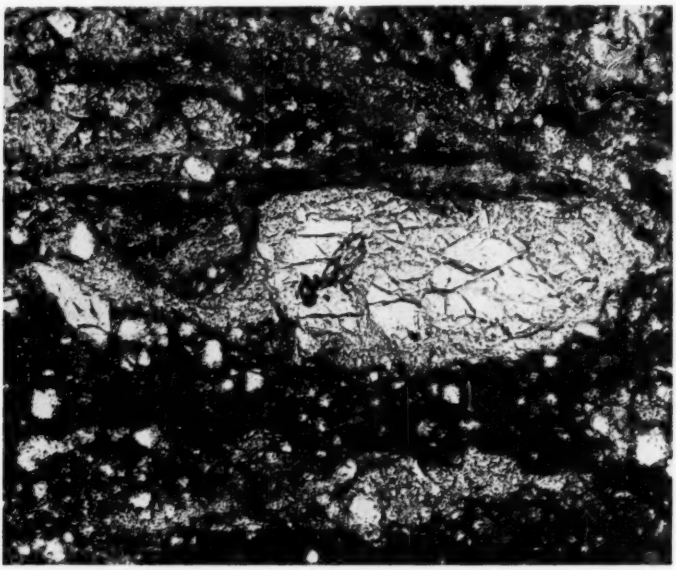

FIG. 9.—Photomicrograph of a feldspar porphyroclast in mylonite. Note the banding produced by difference in granularity of the mylonite groundmass. (Ordinary light, ×60.)

RECURRENCE OF NORTHWEST-SOUTHEAST DEFORMATIONAL MOVEMENTS IN THE NORTHERN CASCADES

It is a curious but well-established fact that the axes of the folds in the late Tertiary rocks of the Cascade Range in Washington run in a general northwest-southeast direction transverse to the main topographic axis of the range. This relationship was first noted by George Otis Smith[1] in the Ellensburg-Yakima region were long,

[1] "Anticlinal Mountain Ridges in Central Washington," *Jour. Geol.*, Vol. XI (1903), pp. 166–77.

moderately dissected anticlinal ridges in the Columbia River lavas and associated sedimentary rocks are conspicuous features of the topography. Transverse folding was also noted by Weaver[1] in the Puget Sound region and by Willis,[2] Calkins,[3] and others in various parts of the Cascade Range.

In the southern part of the Chelan quadrangle three distinct epochs of deformation following northwest-southeast trends occurred in Tertiary time. The first of these took place in the Eocene soon after the deposition of the Swauk formation. The Wenatchee Valley is floored with highly folded Swauk strata whose individual folds comprise parts of a broad synclinorium lying between the Entiat Mountains and the Wenatchee Mountains. The folds within the valley are generally open, but the dips are steep, ranging in most places from 30° to vertical. Along the base of the Entiat Mountains between Cashmere and Eagle Creek the Swauk strata are bowed up abruptly toward the Entiat Range at angles ranging from 55° to vertical. The folding of the Swauk in the Chelan area appears to have been much more severe than in the Mount Stuart district to the south.[4]

Subsequent to the folding of the Swauk formation, but before the outpouring of the Columbia River lavas, the area was cut by a series of normal faults which strike northwest-southeast. Two of these faults define the graben which separates the Swakane gneiss in the Chelan quadrangle into two distinct belts (Fig. 1). Although the faults trend northwest-southeast, they do not always exactly coincide in direction with the axes of the folds in the Swauk. The truncation of a syncline by the southwestern-bounding fault of the graben is clearly visible at Olallie Canyon, and the strike of the strata on the ridge west of Warm Springs Canyon is discordant with the strike of the bounding fault at this locality, although at the actual fault surface the strata have been conformably bowed up for a

[1] Charles E. Weaver, "The Tertiary Formations of Western Washington," *Wash. Geol. Surv. Bull. 13* (1916).

[2] Bailey Willis, "Physiography and Deformation of the Wenatchee-Chelan District, Cascade Range," *U.S. Geol. Surv. Prof. Paper 19* (1903).

[3] F. C. Calkins, "Geology and Water Resources of a Portion of East Central Washington," *U.S. Geol. Surv. Water Supply Paper 118* (1905).

[4] *U.S. Geol. Surv. Folio 106* (1904).

few yards due to drag. The principal normal faults have displacements of at least 1,000 feet at most localities. A minimum figure for the throw of the western-bounding fault of the graben based upon the position of the base of the Swauk formation on both sides of the fault is 4,000 feet, and, if the Swauk formation is of great thickness in the graben, as appears likely, the throw is much more than this figure.

The third episode of deformation in the Tertiary occurred after the outpouring of the Columbia River lavas, and is represented by the formation of slightly dissected anticlinal ridges similar to those described by Smith in the Ellensburg region. In the Chelan quadrangle the Badger-Entiat Mountains are the best example of this folding. As in the districts farther south, the anticlinal ridges are steeper on their northern flank than on the southern.

Study of the Swakane gneiss and the other pre-Tertiary rocks of the Chelan area reveals that these Tertiary deformations were preceded by several periods of orogenic activity which followed the prevailing northwest-southeast trends. The foliation in the Swakane gneiss trends northwest-southeast, and the production of this foliation undoubtedly accompanied a period of orogenic activity in pre-Ordovician time. The pegmatitic injections into the Swakane gneiss lie along a linear belt with the same trend. The mechanical movements which gave rise to the cataclastic effects in the gneiss produced thrusts which strike northwest-southeast and dip to the southwest; and, finally, the longer axis of the Chelan batholith is also concordant with these trends.

The several distinct periods of batholithic activity testify to the occurrence of considerable pre-Tertiary orogenic activity in the Northern Cascades. In the Chelan quadrangle three periods of plutonic invasion are noted. The oldest of these, represented by the granodiorite and quartz-diorite gneiss facies of the Swakane gneiss, is of pre-Ordovician age. The remaining two, the pegmatitic intrusions, which are regarded as the surface manifestations of a batholith not yet fully revealed by erosion, and the Chelan batholith, can only be dated as pre-Swauk formation (Lower Eocene) and post-Swakane gneiss (pre-Ordovician). The large undeformed plutonic masses of the Northern Cascades, such as the Mount Stuart, Index,

and Chelan batholiths, have generally been referred to the late Jurassic, but there is no basis for this assignment other than general petrogenetic similarity to the Sierra Nevada and Coast Range intrusives. That the Chelan batholith and the pegmatite injections are separated in time is proved by the fact that an important episode of mechanical movement occurred between them.

Taking into account the Snoqualmie batholith[1] of Miocene age which lies in the high Cascades to the west, we thus have evidence of four periods of plutonic invasion in this portion of central Washington: one in the Tertiary and three in the pre-Tertiary.

[1] George Otis Smith and F. C. Calkins, "The Snoqualmie Folio," *U.S. Geol. Surv. Folio 139* (1906).

THE GEOLOGY OF BALD BUTTE RIDGE, WASHINGTON

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ABSTRACT

Bald Butte Ridge is a stock of the Idaho batholith which protrudes through and above the late Tertiary basalts. Its rocks form a petrographic series which has crystallized in the following order: hornblende-biotite-tonalite, biotite-tonalite, aplite dikes, quartz-gabbro dikes, pegmatite dikes, and quartz dikes. Most of the rock in the area is biotite-tonalite. It contains schlieren of the hornblende-biotite-tonalite and is cut by all the dikes named above. The tonalites have a poorly developed gneissic structure which was produced by movement in the mass during the early stages of crystallization.

INTRODUCTION

Southeastern Washington is a rolling plain which slopes gently westward from an elevation at its eastern edge of about 2,750 feet above sea level. This plain is broken by several isolated peaks and ridges which rise above it from 700 to 1,100 feet, the most prominent of which are Steptoe Butte, Kamiak Mountain, and Bald Butte.

Bald Butte is near the western end of a series of hills which roughly form a ridge elongated in an east-west direction. The highest point is 3,452 feet above sea level and about 700 feet above the plain. The ridge is approximately 6 miles long and extends from Johnson eastward into Idaho. Its slopes are covered largely with soil formed by the disintegration of the underlying rock, leaving exposed a few large outcrops and many small ones which are well scattered. Trees and shrubs are not abundant, but most of those that do occur are found in the ravines. The area is utilized for agriculture and grazing.

There are no permanent streams in the valleys and ravines of Bald Butte Ridge. The intermittent drainage finds its way into Union Flat Creek on the south, and the South Fork of the Palouse River on the north, both of which are branches of the Snake.

During the past two years the area was used by the writer as a field laboratory for geology classes of the State College of Washing-

[1] The writer is indebted to Professor Albert Johannsen for valuable criticisms, and to Mr. William C. Chun who made some of the thin sections.

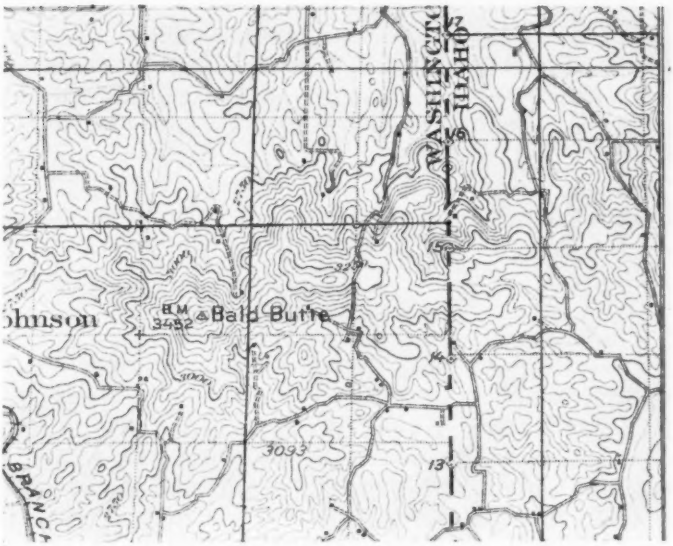

FIG. 1.—Topography of the Bald Butte area. (Published by permission of the Director of the U.S. Geological Survey.)

FIG. 2.—View of Bald Butte from the north

ton, and was visited by him many times. The locality offered the more advanced students an opportunity for field research unaided by published accounts. The only report referring to Bald Butte is one by Treasher,[1] who speaks of the rocks as granite, but makes no mention of an igneous rock series.

GENERAL GEOLOGY

Most of southeastern Washington is underlain by lava flows, basaltic in character, and nearly horizontally bedded. Deep canyons have been cut into these layers, the most impressive being that of the Snake River which has laid bare a vertical section 2,000 feet thick. If, by projection, the original surface of these flows were reconstructed, it would be found to have a gentle westerly dip of approximately 20 feet to the mile. Its elevation surrounding Bald Butte would be about 2,500 feet above sea level and 950 feet below the peak.

The prebasalt topography was one of high relief. The underlying igneous and sedimentary rocks were carved into moderately high mountains separated by deep canyons. Portions of the region, for example, Bald Butte, Kamiak, and Steptoe, were not covered by the flows. Other peaks, of which Granite Point in the Snake Canyon near Wawawai is the best representative, have been uncovered by erosion. The vertical distance between Bald Butte and Granite Point is about 2,700 feet.

A rough estimate can be made of the maximum thickness of the flows in the southeastern portion of the state. When the lava was extruded it filled all of the old depressions up to the present level. The Snake has since cut through the upper 2,000 feet, and in the Bald Butte region Granite Point is the only remnant of the underlying rock that has been exposed. It would perhaps be quite safe to say that to the westward, in some of the deep prebasalt canyons, the base is at least 1,500 feet below the level of the Snake River, making the maximum thickness more than 3,500 feet.

Overlying the basalt is the Palouse formation. Throughout this area it appears to be composed largely of loess with a maximum thickness of about 250 feet. It forms rolling hills and isolated round-

[1] R. C. Treasher, "Geology of the Pullman Quadrangle," Master's thesis (1925), Washington State College, p. 18. Unpublished.

ed knobs, most of which are from 50 to 150 feet high. A very large number of these hilltops have approximately the same elevation. A plane resting on them would have a slight westward dip nearly parallel to the basalt surface and about 250 feet above it.

The major historical events are briefly recorded as follows: (1) the deposition of the early Paleozoic sedimentaries; (2) the intrusion of the Jurassic igneous rocks; (3) a long erosion period; (4) a tremendous outpouring of late Tertiary lavas; (5) deposition of a blanket of Pleistocene loess.

The oldest rocks are the quartzites of Kamiak, Steptoe, and Ringold Hills, and Bald Butte Ridge. These are isolated masses protruding through the basalts and perhaps a part of the metamorphosed sedimentary series, found in the northern part of the state, which is overlapped from the south by the basalts. Some fossil collections have been made in Stephens County at various times during the past several years by Dr. R. H. Jones, Dr. H. E. Culver, Dr. C. C. Branson, and Mr. W. G. Bennett. Branson[1] placed most of the fossils in the early Paleozoic, but stated that some of those collected were as young as Carboniferous.

What happened from Paleozoic times until Jurassic is not as yet known. Certainly some mountain-building occurred. This is indicated in the northeastern part of the state by the high degree of metamorphism and the great amount of folding and faulting the Paleozoic beds have undergone while the intruded Jurassic rocks and overlying basalts have been only slightly disturbed. Late Jurassic dates the intrusion of the Idaho batholith.[2] The igneous inliers in western Idaho and eastern Washington are outlying stocks or bosses of this great intrusion which now is exposed over much of northern Idaho and southern British Columbia. In general, the mass is granodiorite, but locally it grades into tonalite, monzonite, and diorite. These rocks have been cut by dikes both more basic and more acidic.

[1] C. C. Branson, "New Paleontologic Evidence on the Age of the Metamorphic Series of Northeastern Washington," *Sci.*, Vol. LXXIV, No. 1907 (July 17, 1931), p. 70.

[2] Clyde P. Ross, "Mesozoic and Tertiary Granitic Rocks in Idaho," *Jour. Geol.*, Vol. XXXVI (1920), p. 673. Alfred L. Anderson, "Geology and Mineral Resources of Eastern Cassia County, Idaho," *Idaho Bur. of Mines and Geol. Bull., No. 14*, (1931), p. 55, assigns a late Cretaceous or early Eocene age for the Idaho batholith, but his evidence is inconclusive.

At Granite Point horizontally bedded basalts overlie the coarse-grained igneous rocks. Since basalts are extrusive and these coarse-grained rocks intrusive, that portion of the earth's crust which was intruded was eroded before the occurrence of the flows. This hiatus is also noted, although not quite so clearly, at Bald Butte, with the difference that a portion of the intrusive rocks extended above the flows and never was covered by them.

When the lavas were extruded in late Tertiary times, they filled all of the topographic depressions, leaving the higher elevations standing as islands in a sea of basalt. This series of flows was interrupted a number of times. Several interbasalt layers have been found. These are soil beds, or fine-textured tuff, or a mixture of the two, and now and then a poorly indurated sandstone or shale. One which has been studied more than the rest is the Latah formation,[1] which is well exposed near Spokane, Washington. Fossil leaves collected from this formation have been correlated as Miocene.[2] In a number of other places in the basalt, or between the layers, fossil leaves and silicified tree stumps have been found, but very few of them have been described.[3]

The youngest formation is the Palouse loess of Pleistocene age.

BALD BUTTE AREA

The prebasalt rocks, as indicated on the accompanying map, are exposed over an area roughly elliptical in shape and approximately 6 miles east-west and $2\frac{1}{2}$ miles north-south. Only that area was included where igneous rock exposures were found. South of the ridge there are a number of hills which are higher than the Palouse loess and very probably underlain by granitic rocks, although none was seen at the surface. They were therefore not marked as part of the described area even though the soil contained many pieces of rock

[1] J. T. Pardee and Kirk Bryan, "Geology of the Latah Formation," *U.S. Geol. Surv. Prof. Paper No. 140* (1925), p. 1.

[2] F. H. Knowlton, "Flora of the Latah Formation," *U.S. Geol. Surv. Prof. Paper No. 140* (1925), pp. 17–120; E. W. Berry, "A Revision of the Flora of the Latah Formation," *U.S. Geol. Surv. Prof. Paper No. 154-H* (1929), p. 234.

[3] Some fossil leaves from a basalt section in Moses Coulee were collected during the summer of 1931 by Arnold D. Hoffman, of the University of Chicago. The age assigned to the leaves is upper Miocene (*Jour. Geol.*, in press).

from quartz and pegmatite dikes. The tonalites weather readily, as was noted in one hillside which had been exposed by man. Here the biotite-tonalite, with some inclusions of hornblende-biotite-tonalite, is disintegrated to a depth of at least 12 feet.

Two miles northwest from the bench mark on Bald Butte, in the road-cut on the highway from Johnson to Pullman, is an outcrop of a pegmatite dike about 2 feet wide. This is shown on the map as separated from the main mass, because it could only be traced a

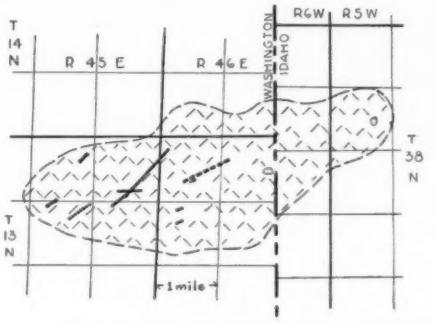

FIG. 3.—Areal geology map
Quartzite—small dotted areas
Biotite-tonalite—marked by diagonal lines
Aplite dikes—heavy broken lines
Pegmatite dikes—heavy dotted lines
Quartz dikes—heavy solid lines
The thickness of the smaller dikes is somewhat exaggerated

short distance from the road. It is, nevertheless, considered as being associated with the Bald Butte series because it is similar to the pegmatites found farther east, and because of its proximity to the described area. Also, it occurs at an elevation above the highest level at which the basalt is found. Between this outcrop and the main exposures, the surface covering contains many pieces of granitic rocks.

At two small localities in the eastern part of the area, some weathered pieces of a quartzitic sandstone were found. These appeared to be similar to the rock protruding through and above the basalt in Kamiak and Ringold Hills about 20 miles to the north-

ward. They were therefore regarded as pieces of float from the sedimentary cover into which the igneous mass was intruded. Some of the quartzitic specimens were cut by a number of very thin dikes, most of which were quartz, the remaining few being quartz-gabbro. It is perhaps due to the metamorphism resulting from these intrusions that float remnants remain.

The main body of the Bald Butte rocks is a medium- to coarse-grained biotite-tonalite which contains schlieren of a medium- to fine-grained hornblende-biotite-tonalite, and which is cut by several series of dikes.

The schlieren are in the form of small, irregular masses, sheets, and stringers. The masses are a foot to several feet across, while the sheets and stringers have a thickness usually of a few inches to a foot or two, and a length of a few feet to 15 or 20. The elongate inclusions are gneissic with their schistosity parallel to the moderately developed gneissic texture of the host rock. This texture is more distinct in some localities than in others. In thin sections the parallel arrangement of the crystals could be seen in the schlieren, but barely a suggestion of it was noticeable in the host rock.

The orientation of the gneissic bands is northeast-southwest. Most of those that were measured were approximately N. 50 E.

In the central and south-central parts of the area, the main tonalite mass is cut by several relatively large masses of light gray, medium-grained aplite. These are in the form of dikes and segregations. The central exposure is a large dike about 400 feet wide which can be traced for about a mile in a direction roughly N. 60 E. Although this intrusion is elongate and wall-like in general form, it is not of constant width, but, instead, thickens and thins. In places its contact is sharp and straight, and elsewhere it is indistinct and seems to merge into the intruded rock.

The exposures of aplite in the south-central portion of the area are lens-like in shape, being from 20 to 50 feet long and from 6 to 10 feet thick. There is no sharp demarcation between them and the surrounding rock.

A few other small areas of aplites were found. The locations of some of them are shown on the geologic map.

Throughout the area there are small, scattered outcrops of a dark

green, fine-grained quartz-gabbro. Exposures are not over a foot or two in width and several feet long. The largest one found had a thickness of $2\frac{1}{2}$ feet and could be traced for about 60 feet. These dike-like intrusions are younger than the aplites. One exposure, although not very distinct, appears to have intruded the large aplite dike, while the quartz-gabbro itself has in turn been intruded by thin, quartz dikes. In other localities this greenish rock has been intruded by both quartz and pegmatite dikes. Those with quartz dikes are the more common.

The last two sets of intrusions are the pegmatite and quartz dikes. Their separation in time cannot be sharply made, since they are really two phases of the same intrusion, the quartz dikes being the more acidic and perhaps a slightly younger phase. A few instances were noted where pegmatite dikes were cut by quartz dikes. A number of examples were seen where a quartz dike was bounded on either side by parallel pegmatite dikes, the dividing line between them being fairly sharp. Several appeared composite, where a pegmatite intrusion had been split and intruded by a quartz dike. In other instances the dividing line between the quartz and pegmatite was not sharply marked. The two phases appeared to have been derived from a single intrusion which had segregated so that the feldspar and muscovite with some quartz crystallized along the walls while the interior separated mainly as quartz.

Most of the dikes are quite small, being only a few inches to a foot in width and exposed for distances of 5–20 feet. Many of them were seen throughout the area, but the largest number was found about a mile northeast of Bald Butte. Some intersect, in which case they appear to have been intruded simultaneously. They strike in all directions, but the northeast-southwest dikes are far more numerous than any others. Where the dip could be determined it was found to be to the south and southeast at various angles from 30° to nearly vertical.

Several large pegmatite and quartz dikes were also found. All but one were directed northeast-southwest, conforming closely to the direction of the gneissic banding.

The largest quartz dike was found at the top of Bald Butte, the highest point in the area, and extending from it northeastward for

about a mile and southwestward for about $\frac{1}{3}$ mile. Close to the peak of the hill another quartz dike cut across the first one in an east-west direction. The east-west dike is the smaller of the two, being about 50 feet wide and $\frac{1}{4}$ mile long. The one directed northeast-southwest is by far the larger and is also the largest in the area. It varies in width from 200 to 300 feet for most of its length. The ends thin out and disappear beneath the soil cover. This dike grades into pegmatite along its margins, beyond which is the tonalite. The pegmatite margins are about 5 feet thick. It is perhaps due to the presence of these dikes that Bald Butte is the highest point in the locality.

There is a quartz outcrop about a mile northwest from Bald Butte. It is elongated northeast-southwest for about 400 feet, and is approximately 60 feet wide. It seems to be composed of at least 2 or 3 parallel dikes rather than one large one.

A pegmatite dike 8 feet wide parallels the southeastern margin of the aplite intrusion in the central part of the area. It is exposed for a distance of but 100 feet.

Southwest of Bald Butte a large pegmatite dike could be traced for about $\frac{1}{2}$ mile. It varies in width from 6 to 12 feet and extends northeast-southwest, N. 45 E.

ROCK DESCRIPTIONS

HORNBLENDE-BIOTITE-TONALITE

This rock is found only in schlieren which range in size from small masses but a few inches in diameter to large masses several feet across and 30 feet long. It is also found in sheets and stringers 15–20 feet long. Samples collected at various points throughout the area vary somewhat in color and grain size, but, in general, it is a dark gray, medium- to fine-grained, gneissic rock. Microscopically it is hypidiomorphic-granular and shows some banding. It is but slightly stained with iron oxide. An average mode is as follows: andesine, 56 per cent; quartz, 28 per cent; biotite, 8 per cent; hornblende, 7 per cent. Present as accessory minerals are a few scattered grains of magnetite and zircon, some small crystals of titanite in the biotite, and many small grains and needles of apatite.

The andesine is rather clear and fresh and is but slightly altered to kaolin and white mica. It also contains minute inclusions of dark

specks and needles, which are perhaps magnetite and titaniferous mica.

The quartz possesses many minute inclusions, most of which look like very small gas and liquid bubbles. The quartz has undulose extinction indicating that the rock has been subject to strain.

The biotite is deep brown and strongly pleochroic. Grains and crystals of titanite are included within it. The hornblende is closely

FIG. 4.—Flowage structure in biotite-tonalite. The darker patches are inclusions of hornblende-biotite-tonalite.

associated with the biotite, and both minerals are arranged roughly in bands which bring out the gneissic texture. The hornblende is deep grass-green in color, strongly pleochroic, and contains some pleochroic halos around zircon.

BIOTITE-TONALITE

This rock forms the main mass in the Bald Butte area, and acts as the host rock for all of the others. It is medium- to coarse-grained and varies in color from light to dark gray. Most of it is light gray with streaks of black biotite. Banding only appears megascopically

and is not noticeable in the thin sections. Microscopically the rock is coarse-grained hypidiomorphic-granular.

The mineral composition is essentially andesine, quartz, and biotite. The andesine ranges between 60 and 72 per cent, the quartz between 12 and 32 per cent, and the biotite between 5 and 11 per cent. An average mode of this rock is as follows: andesine, 65 per cent; quartz, 24 per cent; biotite, 9 per cent; muscovite, 1 per cent. The accessory minerals are apatite, zircon, magnetite, rutile, and titanite.

The andesine is moderately fresh, but is altered along the cleavage and fracture planes where some white mica, kaolin, and epidote have been developed. Some of the crystals of andesine contain many small rectangular inclusions of glass.

The quartz is fresh and clear and contains many grains of the accessory minerals. Much of the quartz shows undulatory extinction, and some of it has cataclastic texture. The rock strain is also shown in the fracturing of the feldspar.

The biotite is strongly pleochroic, dark brown to light brown and yellowish green. Some it is altered to chlorite. The rutile is found as small needle-like blades along the cleavage planes of the biotite. The muscovite, which varies from $\frac{1}{4}$ per cent to 1 per cent, is closely associated with the biotite. Some of the mica is bent.

APLITE DIKES

The aplite is a light gray to white, medium-grained rock which forms one large dike and a number of segregations. Along the margins these aplite masses are found in many instances to grade into the biotite-tonalite.

Microscopically it is hypidiomorphic-granular and ranges in composition between alaskite and granite. An average mode is as follows: microcline, 36 per cent; orthoclase, 12 per cent; oligoclase, 12 per cent; quartz, 28 per cent; micropegmatite, 11 per cent. Some small flakes of muscovite and biotite were present in the hand samples but were lost during the thin sectioning. The accessory minerals are zircon, magnetite, apatite, and a few laths of tourmaline. The tourmaline has good crystal outline and is probably pneumatolytic.

Sagenitic surface texture on the magnetite indicates the presence of titanium.

Small elongated elliptically shaped grains of quartz are included in the microcline, parallel to one direction of the grating texture. The feldspars are somewhat altered to white mica and kaolin. The quartz has been strained until it shows undulose extinction and granular texture.

QUARTZ-GABBRO

The quartz-gabbro is found in small elongated exposures which appear as dike outcrops. This rock is medium- to fine-grained, of granular texture, and varies in color from a light bluish green to a very dark green. Microscopically it is allotriomorphic-granular. On account of its being intruded by many small quartz dikes its mineral percentages vary considerably. The augite ranges between 10 and 50 per cent, and the bytownite between 10 and 41 per cent. Quartz may be absent, or present in amounts up to 60 per cent. A fair average mode appears to be quartz 40 per cent, augite 42 per cent, bytownite 17 per cent. The accessories are a few grains of zircon, some small needles of apatite, many small grains of magnetite, and numerous rounded grains of titanite. In one intsance the titanite totaled 1 per cent.

Throughout, the quartz is quite fresh, but is strained and exhibits undulose extinction. Some partially resorbed plagioclase was found included within the quartz, indicating that most of the quartz, at least, is younger and very probably a later introduction into the rock, the intermixing occurring with the intrusion of the quartz dikes.

The greatest amount of alteration was found in the augite and bytownite, and this varied in different specimens. In the larger number of instances, the augite was partially altered to uralitic hornblende, some shredded biotite, and a small amount of chlorite, the alteration starting principally along the cleavage planes and working inward. In these same cases the bytownite was altered, to a slight extent, to white mica, probably paragonite, along the fractures and the contacts with adjoining minerals. Both albite- and pericline-twinning were noted in the bytownite.

In a specimen collected about a mile west of the top of Bald Butte the augite was partially altered to epidote and uralitic hornblende. The rock carried 48 per cent augite and about one-eighth of it was altered to epidote and hornblende. The bytownite alteration was quite distinctive. Some of it was altered to zoisite and calcite. The percentage composition of this rock is as follows: augite, 39 per cent; epidote, 5 per cent; uralitic hornblende, 1 per cent; bytownite, 41 per cent; zoisite, 5 per cent; calcite, ½ per cent; corundum, 7 per cent; titanite, 1 per cent. No quartz is present.

A mile and a quarter southeast of Bald Butte, in the roadway, there were found close together a few small dark-colored dikes whose mineral composition differs somewhat from any of the others in the area. They are just a few inches in thickness and several feet long. The rock could not be classified, because it varied greatly within short distances. Some of the specimens examined carried as much as 25 per cent of magnetite and some up to 6 per cent of titanite. Green augite was abundantly present, especially in the darker-colored samples. The feldspars associated with these minerals are microcline, orthoclase, and oligoclase. No quartz was noted. The accessories are apatite and zircon. The minerals are quite fresh and show very little alteration. A little sericite has been developed and some of the feldspars have been stained with iron oxide along the fractures.

One feature of mineral associations presents itself. In the tonalites, with or without hornblende, biotite is never absent, but is present in considerable amounts. On the other hand, these gabbroic rocks are always quite free from primary biotite. The rock referred to above is also without biotite, and the presence of augite, magnetite, and titanite classifies it closely with the basic dike rocks. It appears to be a local, but rather more basic intrusion than the quartz-gabbro, which had become associated with potash and acid plagioclase feldspars by a later intrusion.

PEGMATITE AND QUARTZ DIKES

The pegmatite dikes are medium- to very coarse-grained rocks. The minerals are quartz, orthoclase, microcline and muscovite. Since the grain is usually very coarse, many of the feldspar crystals being 2 inches long, a percentage estimate from a study of the thin

sections would be of no value. In spots the rock is mostly quartz. Other samples nearby are mostly orthoclase, and locally small books of muscovite are seen. Some of the quartz crystals appear striated as a result of the development of long, narrow, parallel crystal faces, a number of which have impressed their form on the adjacent feldspar crystals.

The quartz dikes are almost monomineralic, only about $\frac{1}{2}$ per cent of the rock being made up of small flakes of muscovite and biotite. A few garnets were found in or close to several of the pegmatitic segregations. Most of the quartz is strained and shows undulatory extinction. At the very eastern edge of the area some quartz crystals about $\frac{1}{4}$ inch long were found in a small drusy cavity in a quartz dike. Many of them were pyramids, but no regular arrangement of the crystals could be noted.

PETROLOGY

The petrologic history of these igneous rocks, for the purposes of this report, begins with the intrusion of the magma into the Bald Butte area. The chemical composition as shown by the minerals makes it apparent that the magma was but a differentiate of a larger one. Since the Bald Butte rocks occur as an arm or stock of the Idaho batholith, the parent magma was most likely that of the latter. The discussion here is confined to the origin of the rocks from the local magma, their structures, and textures.

The Bald Butte magma was in motion during the earlier stages of its crystallization. This is shown by the gneissic textures in the schlieren, the roughly developed gneissic texture in the biotite-tonalite, and the parallel arrangement of most of the elongated schlieren. That the parallel arrangement of minerals and inclusions occurred at this time and not after the mass had completely solidified, is shown by the small amount of disturbance seen in the dikes which are younger than the biotite-tonalite. The quartz in the younger rocks, which shows undulatory extinction and a small amount of cataclastic texture, indicates the strain to which they have been subjected. Such strain may have been produced by pressures resulting from shrinkage caused by cooling, or to settling, or to a very slight regional warping.

In the early stages of the intrusion the outer layers in their new

surroundings were in contact with rocks considerably colder than the magma. Rapid crystallization thereby ensued, making the outer rock medium- to fine-grained. Most of the hornblende and biotite crystals oriented themselves approximately normal to the pressure caused by the intrusive forces. Further movement of the mass permitted the inner, more coarsely grained rock, while still in a very plastic state, to tear off masses of the earlier-formed outer layer. Beyond this stage there seems to have been very little movement. Shrinkage produced some fracturing, and after the final crystallization enough pressure was exerted to strain the quartz.

The gneissic banding is oriented northeast-southwest. Where the dip of some of the pegmatite dikes could be seen it was to the south and southeast at angles ranging from 30° to nearly vertical. These facts would indicate an overthrusting pressure from the southeast, or an underthrusting pressure from the northwest. Since the area is a stock of the Idaho batholith to the eastward, a thrust from the southeast seems to be more reasonable.

Bald Butte Ridge, as has already been indicated above, is largely a biotite-tonalite, which contains many schlieren of an older hornblende-biotite-tonalite, and which has been intruded by aplite dikes, quartz-gabbro dikes, pegmatite dikes, and quartz dikes in the order named. Looking at the rocks from a quantitative basis, the biotite-tonalite forms most of the intrusive mass. All of the other rocks combined make up but a very small percentage of the total. In the order of decreasing volumetric importance they are as follows: the quartz dikes, the pegmatite dikes, the aplite dikes, the hornblende-biotite-tonalite schlieren, and the quartz-gabbro dikes. A study of the relative proportions of these rocks and the composition of the minerals forming them should show approximately the chemical composition of the magma from which they were crystallized. Such a study brought out certain variations from what is usually considered the normal type. The principal differences were relatively low potassium and sodium and relatively high calcium and silica.

The low percentage of potassium is indicated by the presence of but very small amounts of the potassium-bearing minerals—orthoclase, microcline, and muscovite. The biotite-tonalite, which forms the main mass of the rock, carries no potash feldspar and only small

quantities of muscovite. The largest amount of muscovite seen did not exceed 1 per cent. The aplite and pegmatite dikes do contain large percentages of orthoclase and microcline, but they comprise only a very small part of the total mass.

The high percentage of silica is shown by the abundance of quartz, with a few minor exceptions, in all of the rocks in the area. The relatively high percentage of calcium is suggested by the types of plagioclase feldspar present. Andesine occurs in the tonalites and bytownite in the quartz-gabbro, both being more calcic than those usually found in such highly acidic rocks.

The differentiation which produced the Bald Butte rocks probably had a course somewhat as follows: When the magma was intruded into rocks of much lower temperature than that necessary to keep it in liquid form, rapid crystallization of the outer portion developed the fine- to medium-grained hornblende-biotite-tonalite. The rate of crystallization prevented differentiation and the chemical composition of this rock is very likely close to the composition of intruded magma. The solid outer layer retarded the escape of heat and so permitted slow cooling of the inner mass. The early-formed augite, basic plagioclase, magnetite, ilmenite, and titanite sank to the lower regions of the magma basin. Some of these minerals were resorbed partially or wholly, the resultant being a basic liquid containing crystals of calcic plagioclase and ferromagnesian minerals. The sinking of the more basic constituents from the upper portions left behind a liquid which gradually became more and more acidic. Further crystallization formed rocks which graded from the crustal hornblende-biotite-tonalite inward to biotite-tonalite carrying a small amount of muscovite, to leucogranite and alaskite which formed the aplite dikes. The end-product was the liquid from which the pegmatite and quartz dikes were crystallized.

The change from liquid to solid gradually reduced the volume, and the general loss of temperature brought about shrinkage. While the mass was still quite liquid readjustment was effected principally by flowage. As crystallization progressed the mass changed from a semi-solid to one quite rigid. Readjustment in these later stages produced fracturing.

The first fracturing occurred at a time when the mass was not in-

tensely rigid. The fractures were uneven, irregular, and in some instances much like pockets and cavities. These were soon filled by the aplites in the form of dikes or segregations.

At a still later stage the mass had become so rigid that deep fractures were developed to relieve the stresses set up by the crystallization and cooling. These fractures extended from the outer layers and overlying beds deep into the rock body, and formed avenues through which came the quartz-gabbro fluid. The later stages of this readjustment developed the fractures in which the pegmatite and quartz dikes crystallized.

Upon reaching the upper cooler areas the quartz-gabbro liquid quickly crystallized and formed fine-grained dikes. A very small amount of the quartz associated with the augite and bytownite may have come in with this solution, but certainly the high percentages of quartz are due to the later intrusions of the quartz dikes.

The final liquid was well distributed through the intercrystal spaces of the mass. It moved into the open fractures, the areas of lesser pressure, and much of it was not very far removed from where it had existed before the readjustment. It was a highly siliceous liquid rich in mineralizers most of which had been moved into an environment but slightly changed. That which found its way into the cooler outer regions crystallized at a relatively rapid rate and formed a medium-grained rock. That which did not move far formed coarse-grained rocks. From this liquid the pegmatites were first crystallized, the residual forming the quartz dikes.

FIELD EVIDENCE TO DISTINGUISH OVERTHRUSTING FROM UNDERTHRUSTING[1]

T. S. LOVERING[2]
Golden, Colorado

ABSTRACT

If a thrust fault breaks from an overturned fold in a zone of tear faulting, the movement of the walls of the tear faults shows the direction of movement of the adjacent thrust block and thus indicates whether underthrusting or overthrusting has taken place. Similarly, a marked swing of formations toward or away from the axis of the overturned fold as a thrust fault is approached suggests underthrusting or overthrusting, respectively. Application of these criteria to the Seminoe Mountain thrust fault and to the Williams Range thrust fault indicates that the first is an overthrust and that the second is an underthrust.

No field criterion has been suggested for discriminating between the active and passive blocks of thrust faults in the discussions of overthrusting and underthrusting, as far as the writer is aware. While studying the geology and ore deposits of the Montezuma quadrangle, Colorado, suggestive field evidence was found, and is discussed in this article. The criteria offered do not apply to all thrust faults, but only to those that can be traced into folds or that are cut by tear faults. A large number of thrust faults, however, fall into one of the above categories, and the field determination of the active and passive blocks, wherever possible, should contribute to our knowledge of tectonics.

The term "underthrust fault" implies a low-angle fault in which the foot wall moved forward actively under a passive hanging wall that remained nearly stationary. An "overthrust fault" implies the opposite set of conditions, an active hanging wall moving forward over a passive, relatively stationary, foot wall.

The determination of the active and passive blocks of thrust faults is based on the relative movement in the zone where the thrust fault passes into a fold or breaks from a fold along a tear fault. Tear faults commonly strike almost at right angles to the thrust faults, stand almost vertical, and occur between more active and less active re-

[1] Published by permission of the Director of the U.S. Geol. Surv. and the Col. Metal Mining Fund.

[2] Associate Geologist, U.S. Geological Survey.

gions of folding, where the fault gives way to an overturned fold (see Fig. 1). Thus, tear faults are commonly found in a nearly vertical zone of low-angle shearing, and their steeply dipping walls have moved past each other almost horizontally. They have formed in response to the shearing stresses that caused thrust faulting in the region of active folding and are, therefore, approximately parallel to the direction of the thrust, which is commonly almost at a right angle to the axis of the folds.

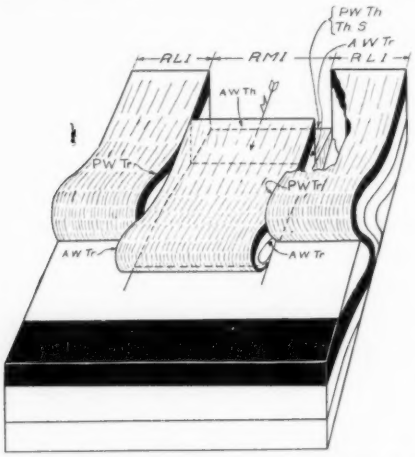

FIG. 1.—The relation of tear faults to overturned folds and to thrust faults that are formed by overthrusting. *R L I*, region of less intense deformation or inactive block; *R M I*, region of more intense deformation; *A W Th*, active wall of thrust fault or active block; *P W Th*, passive wall of thrust fault or passive block; *Th S*, surface of thrust fault; *A W Tr*, active wall of tear fault; *P W Tr*, passive wall of tear fault; *M*, direction of movement of active block.

If, in the region of thrust faulting, one of the fault blocks is decidedly more active than the other, the tear faults that separate the active block from the region of less intense deformation will be stronger and more evident than those between the relatively passive block and the region of less intense deformation (see Fig. 2). If in the region of intense deformation one block remains entirely passive while all the thrusting takes place by movement of an active block, the tear faults and the zone of shearing will be concentrated in the

zone between the active block and the region of less intense deformation, and there will be relatively little shearing between the passive block and the region of less intense deformation. In this type of thrusting, if the relatively passive block were dragged slightly by the active block, the shear zone between the passive block and the region of less intense deformation would show the same direction of movement as that found between the active block and the region of less intense deformation (see Figs. 2 and 3). If only underthrusting or

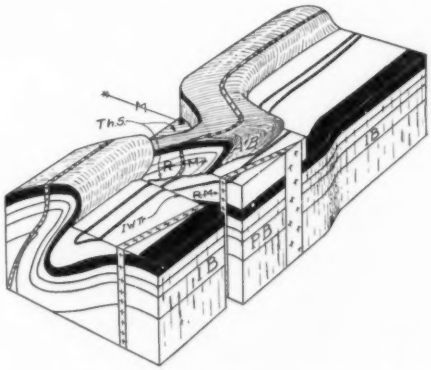

FIG. 2.—Overturned fold passing gradually into overthrust fault at one side of region of more intense deformation and broken by tear fault at other side. *M*, direction of movement of active fault; *R M*, relative movement of active and passive walls of thrust fault with respect to the inactive wall of the tear fault is shown by arrows; *Th S*, thrust fault surface; *I W Tr*, inactive wall of tear fault; *A B*, active or hanging wall block of overthrust fault; *P B*, passive or foot wall block of overthrust fault; *I B*, inactive block or region of less intense deformation.

overthrusting is involved in the formation of a thrust fault, the direction of movement of the walls of the tear fault will indicate which type of movement took place, and this direction of movement should be the same whether the tear faults studied lie between the upper block and the region of less intense folding or between the lower block and the region of less intense folding (see Figs. 2 and 3). If two actively opposed forces were involved in the formation of the thrust fault, one of which caused underthrusting and the other overthrusting, the direction of movement in the tear fault zone between the

upper block and the less active region of deformation would be opposite to that between the lower block and the region of less active deformation (see Fig. 4).

As the zone of tear faulting commonly exposed to view lies between the upper block and the region of less intense deformation, it will be well to consider it specifically. The active wall in a tear fault is the wall next to the region of more intense deformation; the passive wall is the wall on the side next to the region of less intense

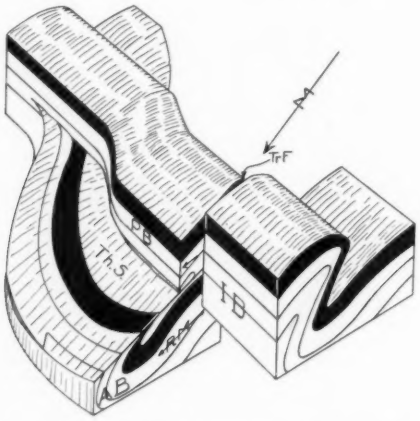

FIG. 3.—Underthrust fault passing into overturned fold at one end and broken by tear fault at the other end. *M*, direction of movement of underthrust fault block; *R M*, relative movement of hanging and foot walls of thrust fault along tear fault is shown by length and direction of arrows; *Tr F*, tear fault; *Th S*, surface of foot wall of thrust fault; *I B*, inactive block of tear fault or region of less intense deformation; *P B*, passive or hanging wall block of underthrust fault; *A B*, active or foot wall block of underthrust fault.

deformation. If the active wall of a tear fault has moved in the direction that the hanging wall of the thrust fault has apparently moved (Fig. 2), it indicates an overthrust fault; on the other hand, if the active wall of the tear fault has moved in an opposite direction, e.g., in the direction that the foot wall of the thrust fault has apparently moved, it indicates an underthrust fault (Fig. 3).

Where tear faults are absent the swing of the formations on the overturned side of the fold where it passes into the thrust fault is

suggestive. If the formations bend toward the axis of the fold as they approach the thrust fault, the bend suggests drag by an underthrust block (Fig. 3). If, on the contrary, they tend to bend away from the axis of the fold, the outward bend suggests the force of an overthrust block crowding them away from the axis of the fold (Fig. 2). As illustrated in Figures 5, 6, 7, and 8, the plan view of a thrust fault that has undergone erosion is dependent on many factors. Upwarping and downwarping of the active region of deforma-

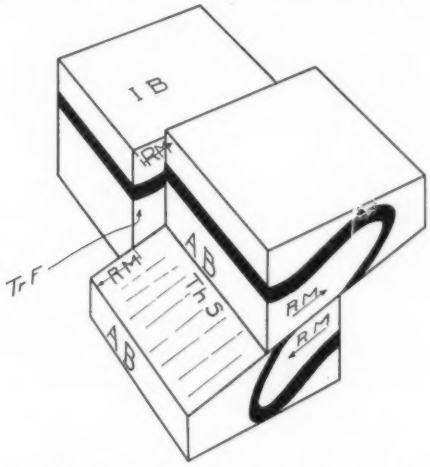

FIG. 4.—Block diagram illustrating relative movement of fault blocks when both overthrusting and underthrusting occur. *I B*, inactive block; *A B*, active block; *Tr F*, tear fault; *Th S*, surface of foot wall of thrust fault; *R M*, relative movement of hanging and foot walls with respect to the inactive block.

tion will shift the outcrop of the thrust fault in opposite directions. If the inclination of the thrust fault is gentler than the dip of the beds that it crosses, it will shift the outcrop of the fault away from the axis of the fold; if it is steeper than the inclination of the beds that it crosses, it shifts the outcrop of the fault closer to the axis of the fold. It should be understood that the block diagrams used as illustrations are idealized and schematic. No attempt has been made to represent the usual complexities of structure that accompany underthrusting or overthrusting.

The application of the foregoing criteria is illustrated in the following discussions of a specific case of overthrust faulting and one of underthrust faulting.

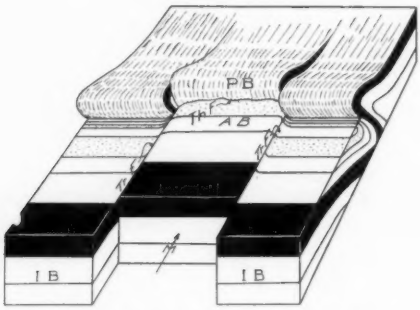

FIG. 5.—Deformation from horizontal movement of underthrust foot wall. *P B*, passive block or hanging wall of underthrust fault; *A B*, active or foot wall block of underthrust fault; *Th F*, trace of thrust fault; *Tr F*, tear fault; *I B*, inactive block or region of less intense deformation.

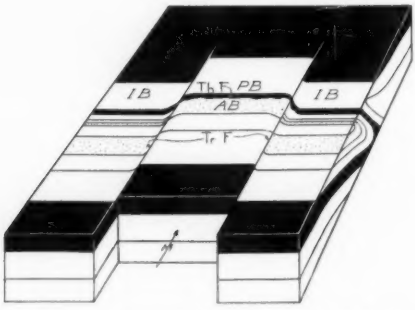

FIG. 6.—Effect of erosion on outcrop of underthrust fault in which the active foot wall moved horizontally, see Figure 5. *I B*, inactive blocks; *P B*, passive block of thrust fault, i.e., the hanging wall; *A B*, active block of thrust fault; *Th F*, trace of thrust fault at surface; *Tr F*, tear fault; *M*, direction of movement of underthrust fault.

THE SEMINOE OVERTHRUST FAULT

The Seminoe thrust fault occurs in south-central Wyoming, about 40 miles northeast of Rawlins.[1] The Pathfinder uplight stretches

[1] T. S. Lovering, "The Rawlins, Shirley, and Seminoe Iron Ore Deposits, Carbon County, Wyoming," *U.S. Geol. Surv. Bull. 811* (1929), pp. 209, 227–29.

west-northwestward 100 miles from Difficulty to Haley and is about 30 miles wide in the vicinity of the Pathfinder reservoir. The south-

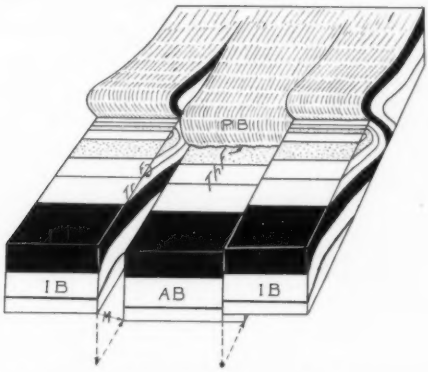

FIG. 7.—An underthrust fault in which downwarping has accompanied the movement of the active foot wall block. *I B*, inactive block or region of less intense deformation; *A B*, active or foot wall block of underthrust fault; *P B*, passive or hanging wall block of underthrust fault; *M*, direction of movement; *Th F*, outcrop of thrust fault; *Tr F*, tear fault.

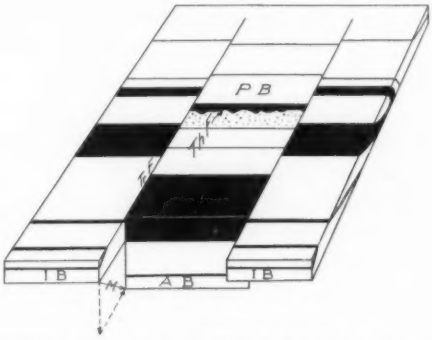

FIG. 8.—Effect of erosion on underthrust fault in which the active block has been warped down, see Figure 7. *I B*, inactive block or region of less intense deformation; *A B*, active or foot wall block of underthrust fault; *P B*, passive or hanging wall block of underthrust fault; *M*, direction of movement; *Th F*, outcrop of underthrust fault; *Tr F*, tear fault.

west border of this arch has buckled and broken at many places, and overturned folds and thrust faults occur in several localities. In the

Ferris and Seminoe Mountains on the south side of the Pathfinder uplight pre-Cambrian formations overlie Paleozoic and Mesozoic sediments (see Fig. 9). Although the structure is complicated by minor faults, the more significant relations of the major fault are apparent. Near the eastern end of the Ferris Mountains the almost vertical Paleozoic sediments swing east and southeast and, coincident with the change in strike, the beds become overturned, dipping toward the mountains at lower and lower angles until they reach the Seminoe fault, the major thrust fault of the region, where they disappear under a mass of pre-Cambrian rocks. The swing in the direction of the beds as they approach the thrust fault suggests the drag of an active hanging wall block as it was thrust southward over a relatively passive foot wall. The eastern limit of the thrust fault is marked by a tear fault striking nearly north and south. The relative direction of movement along this tear fault is clearly indicated by the marked southward drag of the vertical and overturned lower Paleozoic beds bordering the tear fault on the east. They have apparently been pulled southward by the active hanging wall of the thrust fault.

The structure due to the minor faults of the region is less clearly shown. Much of the region southwest of the pre-Cambrian area is covered by windblown sand, and the structure of the sediments is not accurately known. Southeast of the tongue-like mass of overthrust crystalline rocks, the overturned sandstone at the base of the Mesaverde formation lies about 1 miles southwest of the overturned Triassic red beds, although the stratigraphic interval between them is approximately 7,000 feet. It is unlikely that folding alone would thin the beds enough to make them fit this space. A similar discrepancy in the thickness of the section is found to the northwest where the Steele shale, dipping gently southeast, is separated by a mile of windblown sand from almost completely overturned red beds. The stratigraphic distance between the formations is about 2,500 feet, and the synclinal structure between them involves the repetition of some of the beds. The distance between the outcrops seems inadequate for an unfaulted overturned fold. These stratigraphic and structural relations indicate the presence of a concealed northwesterly trending thrust fault in the sediments. The southern half of the

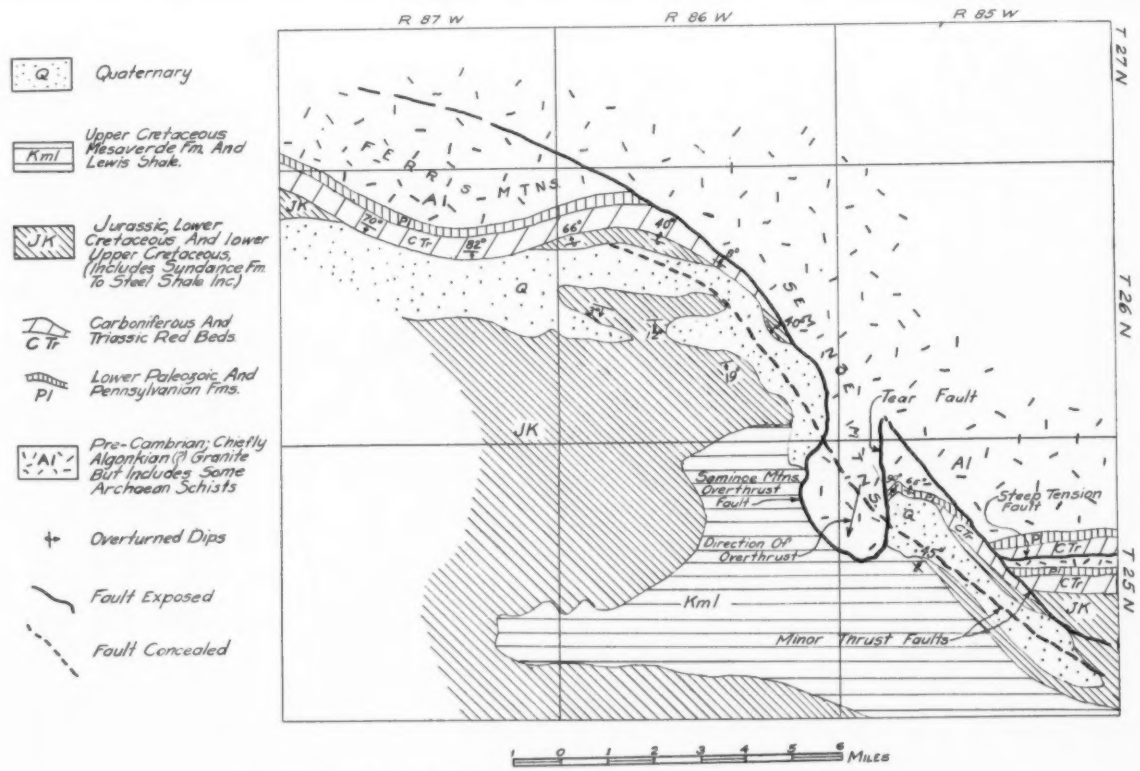
Q Quaternary
Kml Upper Cretaceous Mesaverde Fm. And Lewis Shale.
JK Jurassic, Lower Cretaceous And Lower Upper Cretaceous, (Includes Syndance Fm. To Steel Shale Inc.)
C Tr Carboniferous And Triassic Red Beds
P l Lower Paleozoic And Pennsylvanian Fms.
Al Pre-Cambrian; Chiefly Algonkian (?) Granite But Includes Some Archaean Schists
Overturned Dips
Fault Exposed
Fault Concealed
R 87 W
R 86 W
R 85 W
T 27 N
T 26 N
T 25 N
FERRIS MTNS.
SEMINOE MTNS.
Tear Fault
Seminoe Mtns. Overthrust Fault
Direction Of Overthrust
Steep Tension Fault
Minor Thrust Faults
Miles

FIG. 9.—Sketch map of Seminoe Mountain thrust fault, Carbon County, Wyoming

concealed fault is paralleled about a mile to the northeast by another minor overthrust that brings the pre-Cambrian rocks on the northeast against the lower Paleozoic sediments. Both of these minor overthrusts are believed to have been formed during the early stages of folding and faulting which culminated in the major Seminoe overthrust. The repetition of the steeply dipping sediments east of the minor thrust faults is probably due to a tension fault, as the displacement is the opposite of that on the major and minor thrust faults.

THE WILLIAMS RANGE UNDERTHRUST FAULT

In recent years much thrust faulting has been found in Colorado. These structures are well shown on the new geologic map of the state, which is in preparation by the U.S. Geological Survey. The structural features described below occur on the western side of the Front Range in the central part of Colorado.

The Colorado Front Range is an elongate dome extending north from Canon City, Colorado, to Laramie, Wyoming. Although it is distinctly north-south in its regional trend, it is made up in large part of northwesterly trending folds arranged *en échelon*. These northwesterly flexures are commonly asymmetric, having steep western limbs and gently dipping eastern limbs, and, where deformation was severe, the folds are broken along their axes by strong reverse faults that have dropped the western part of the fold.

Thrust faults have been found on both borders of the range, but the largest one known at present is the Williams Range thrust fault on the western side (see Fig. 10). This fault has been traced from the Breckenridge district 50 miles northwestward to Muddy Butte, a short distance beyond Kremmling. Farther northwest it is hidden under Tertiary sediments, and the structure is unknown. At its southeastern end, in the Breckenridge district, it passes into an overturned fold that continues southeast into South Park, where it is hidden beneath Pleistocene gravel.

A narrow belt of fractures crosses the Front Range almost at right angles to the strike of the northwesterly trending fold and extends northeast from Breckenridge to Boulder. This transverse fissured zone appears at the western edge of the Front Range where the overturned fold breaks and passes into the Williams Range thrust fault.

Here it may be regarded as a zone of tear faulting related to the thrust fault. Nearly all the Tertiary igneous activity in the central part of the Front Range was localized in this strip of fractured

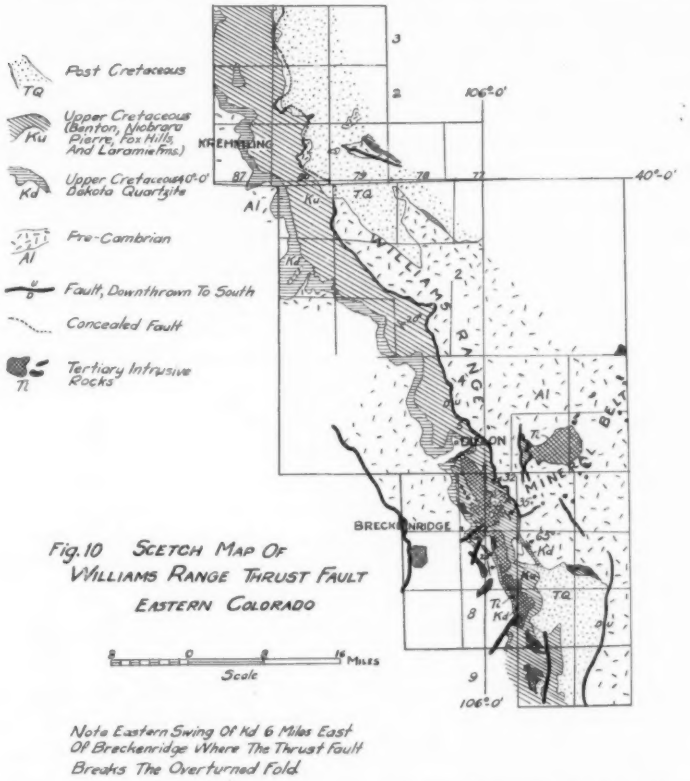

Fig. 10 Sketch Map Of Williams Range Thrust Fault Eastern Colorado

Note Eastern Swing Of Kd 6 Miles East Of Breckenridge Where The Thrust Fault Breaks The Overturned Fold

FIG. 10.—Sketch map of the Williams Range thrust fault in central Colorado. Note the eastward swing of the upper Cretaceous Dakota quartzite 6 miles east of Breckenridge, where the Williams Range thrust fault breaks from the overturned fold.

ground. This concentration of the porphyritic Tertiary intrusives in a narrow zone was early recognized by miners who named it the "porphyry belt." Within the pre-Cambrian terrane of the Front Range the porphyry belt is only a few miles wide, but in the sedi-

ments to the southwest the intrusions spread out over a much greater distance, following a zone of weakness near the trough of the marked syncline that separates the Front Range from the Mosquito–Ten Mile Range. Aside from the ore deposits at Cripple Creek, nearly all the important ore deposits in the Front Range are genetically related to the porphyries and are confined to the same narrow strip, and for this reason the names "porphyry belt" and "mineral belt" are used interchangeably.

Most of the veins in the mineral belt strike from east to northeast but some northwesterly veins are present. The important veins occupy premineral faults of appreciable displacement, and on many of them the direction of the pre-mineral movement has been ascertained. The easterly trending veins in the western half of the Front Range show a marked uniformity in the direction of their displacement. As far as known, in all faults of this system in which the direction of movement has been found, the northern wall moved eastward almost horizontally.[1] Northeasterly and northwesterly trending faults commonly show less horizontal than vertical movement, and both reverse and normal faults are common. The easterly trending faults of the mineral belt show the presence of widespread, horizontal shearing forces. The Williams Range thrust fault dips east-northeast at a low angle and also records the presence of a regional compressive force that was nearly horizontal. The force was less active southeast of Breckenridge where the overturned fold is not broken, but north of this district, where the fold is broken by the thrust fault, there is a horizontal displacement of more than 4 miles.[2] In the transition zone between the overturned fold and the region of more active deformation, strong shearing stresses must have developed, and it is believed that these shearing stresses are recorded in the easterly trending premineral faults of the mineral belt to the northeast.

The direction of movement of these shear faults in the mineral belt

[1] S. H. Ball, J. E. Spurr, and G. H. Gary, "Economic Geology of the Georgetown Quadrangle, Colorado, *U.S. Geol. Surv. Prof. Paper 63* (1908), p. 162; T. S. Lovering, "Geology and Ore Deposits of the Montezuma Quadrangle, Colorado," *U.S. Geol. Surv.* unpublished MS.

[2] T. S. Lovering, "The Williams Range Thrust Fault" (abstract), *Geol. Soc. Amer. Bull.*, Vol. XXXIX (1928), pp. 173–74.

shows that the north side moved east. This is equivalent to saying that the active walls of the tear faults moved in the apparent direction of the movement of the foot wall of the Williams Range thrust fault and in the opposite direction to the apparent movement of the upper or hanging wall of the thrust fault. This type of movement in the tear fault zone strongly suggests an underthrust fault. Moreover, in the region of less intense deformation south of the fault, 6 miles east of Breckenridge, the overturned formations bend eastward toward the axis of the fold as they are traced northward into the shear zone, where they disappear under the hanging wall of the thrust fault. This bend toward the axis of the overturned fold, as the formations are followed into the region of more intense deformation, also strongly suggests underthrust faulting. Since the direction of movement of the shear faults and the swing of the formations both suggest underthrusting, the Williams Range thrust fault with some degree of assurance can be ascribed to underthrusting.

DISCUSSION OF "THE ISOSTASY OF THE UINTA MOUNTAINS,"[1] BY ANDREW C. LAWSON

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Although hesitant to discuss the subject of isostasy, the writer cannot agree with Dr. Lawson's article on "The Isostasy of the Uinta Mountains." Dr. Lawson gives far greater credit to isostasy and assigns much less to compression in the formation and present relief of these mountains than is warranted by the geologic evidence presented by them. The outstanding structural features of the Uinta Mountains, the overturned anticline, and the low angle thrust faults, should prove that compression, not isostasy, was the dominant factor in the diastrophism. The principle of isostasy is a factor which must be considered in mountain building, but only in a minor not a major rôle in the formation of mountains of this type.

Two statements of Dr. Chamberlin's fit this case very aptly:[2]

> To the geologist the recognized geologic facts, which he can understand and appreciate, are vastly more convincing than mathematical interpretations based upon assumptions, some of which he does not understand, and others of which seem to him clearly at variance with the actual earth conditions. Though the mathematical mill grinds with precision, the mill has been fed so commonly with defective assumptions and untruths in past application to geological problems that the grist has often proved more deleterious than beneficial. Past experience justifies the geologist in viewing with caution and skepticism mathematical deductions which seem inconsistent with field observations logically interpreted.
>
> Isostasy, though a working principle, can be overworked.

Dr. Lawson presents a formidable array of figures to sustain his theory, but these figures, precise as they are, are open to question. It would appear that he has based his calculations for isostasy in the Uinta Mountains largely, if not entirely, on the cross sections and findings made by Powell in the late seventies. These would seem to

[1] *Jour. Geol.*, Vol. XXXIX, pp. 264–77.

[2] Rollin T. Chamberlin, "Isostasy from the Geological Point of View," *Jour. Geol.*, Vol. XXXIX, pp. 1 and 17.

be rather insecure foundations, as Powell's interpretation of the Uinta fault has been proved incorrect, later investigations showing the fault or series of faults to be overthrusts, whereas Powell's cross-sections show this fault as vertical with the formations bordering it on both sides as nearly horizontal in attitude. These sections do not show the steeply upturned and overturned attitude of the sedimentary strata bordering the fault on the north, or downthrow side, or the folded conditions to the south of it. A reconstructed

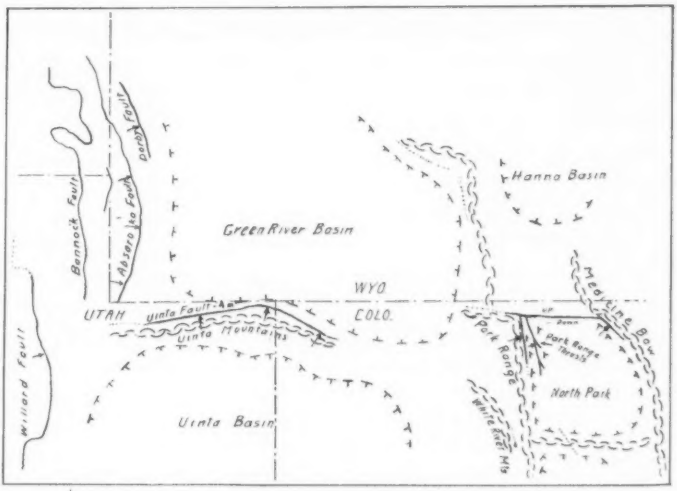

FIG. 1.—Sketch map showing the distribution of the major thrust faults in Colorado, Wyoming, and Utah. Dotted lines are faults that inferred or doubtful.

cross-section of the range, as it would appear had not erosion taken place, would show it to be an overturned and thrust-faulted anticline. The fracturing has occurred in a series of thrust faults ranging from nearly vertical to overthrusts of several miles, 45 degrees being about their average dip. Ver Wiebe[1] says: "The Uinta thrust fault appears to have allowed old Paleozoic sediments to override much younger strata from south to north for a distance of 4 to 5 miles. The dip of this fault is thought to be about 45 degrees and the displacement amounts to as much as 30,000 feet."

[1] W. A. Ver Wiebe, *Oil Fields in the United States*, p. 487.

Dr. Lawson's figures may be questioned still further, at least to some extent. His application of isostasy to the elevation of the Uinta Mountains seems to be based upon a close estimate of the material removed from the range and upon definite elevations credited to these mountains at different periods in their history. Our ideas of the actual conditions which prevailed at these periods are still speculative. Also, the many geologic processes which enable us to trace out the history of a mountain range do not act in a uniform manner throughout its extent. Owing to these uncertainties, the writer does not believe that either the amount of material removed by erosion or

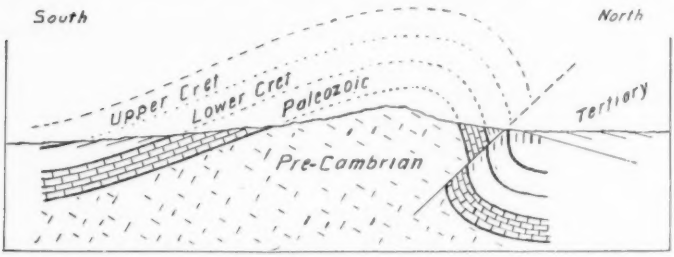

FIG. 2.—A cross-section of the Uinta Mountains, showing the manner in which the anticline is overturned and thrust faulted. The scale and geology are only approximated.

the exact amount of uplift at different stages can be determined with the precision implied in Dr. Lawson's article.

Dr. Lawson states that orogeny due to compression began in the late Cretaceous. His opening statements[1] are:

> The main uplift of the Uinta Mountains is a broad simple anticlinal arch which attained a mean altitude of 1.82 miles in early Tertiary time, as a result of an orogenic upthrust of 0.66 miles added to a general epeirogenic uplift of the region of 1.16 miles. The culminating point of the arch was of course higher than the mean altitude, but probably never exceeded 2.8 miles. The upthrust of 0.66 miles necessitated a downthrust of the sial into the sima of 5.94 miles to maintain isostatic equilibrium. Under erosion the mountain column lost load and rose by flotation. . . . . The rise separated the range from the plateau on the north by an upthrow of 3.98 miles on the north-flank fault.

The statement, "The mountains attained a mean altitude of 1.82 miles by early Tertiary," implies that this elevation was the total elevatory movement by this time, that due to compression plus the

[1] *Op. cit.*, p. 264.

general uplift of the region. The amount of erosion that must have taken place during this uplift is not taken into consideration, unless these figures are meant as the actual mean elevation in regard to sea level. In either case, the amount of uplift that seems to be credited to compression, a mean of 0.66 miles with an extreme of 1.64 (2.8 − 1.16), is entirely too small to effect the folding shown by these mountains. The least measure of elevatory movement which necessarily must result from compressive forces forming and overturning an anticline of this size would be much greater than is assigned to them. As the thickness of sedimentary strata involved in this deformation is 3 miles or more, plus an unknown amount of the pre-Cambrian, this movement should be measured in miles instead of fractions. The relief of these mountains, however, either during or at the climax of this orogeny, is largely a matter of speculation, as erosion must have been active while the fold was rising. It is not very probable that they attained any extreme elevations at any time, i.e., much higher than at present.

Dr. Lawson says that the epeirogenic uplift was independent of the orogenic and later uplifting, and is used by him as a datum plane upon which to base his calculations. This statement is quite true, and his figures for this uplift, (1.16 miles) agree quite closely with those given by Schuchert and others. The writer believes, however, that the general opinion is that the epeirogenic uplift took place at a later date than Dr. Lawson asserts. Schuchert[1] states that this movement began in the Miocene and continued throughout the Pliocene and perhaps, also, the Pleistocene. The preponderance of evidence tends to show that the greater part of the Basin region was low land during much of the Tertiary and that no marked climatic or other changes took place, such as would suggest a widespread elevation of the land until about Pliocene time. Some[2] crustal unrest continued throughout this period however; local folding and thrust faulting took place, probably culminating with the widespread elevatory movements in the Pliocene.

The writer has attempted to point out the fact that the Cretaceous orogeny was much greater than Dr. Lawson states, and that

[1] Pirsson and Schuchert, *Text Book of Geology*, Vol. II (New York: Wiley & Sons, 1924), p. 606.

[2] Ver Wiebe, *op. cit.*, p. 448.

although the exact elevation of the mountains is not known, it was at least greater than Dr. Lawson's figures would infer. Following this period of diastrophism the mountains were probably reduced to a comparatively low relief if not baseleveled by the Tertiary erosion before the next stage of crustal unrest and uplift set in. The movements in this later stage are largely responsible for the present relief of the Uintas, and it is in the re-uplift of these mountains that the theory of isostasy runs most contrary to the structural and geological evidence shown by them. Dr. Lawson assigns the greater part of this uplift (3.98 miles) to isostatic adjustment, due to the loss of load and rise by flotation of the mountain column. This movement is assumed to be vertical, taking place by displacement along a vertical fault on the north and a flexure on the south. The fact that the fault is not vertical but is a thrust fault, dipping on an average 45 degrees south, shows that this movement was not vertical, but on a 45-degree incline. The type of rocks involved in this faulting and folding, massive limestones, sandstones, and quartzites, would suggest that the Uinta fold was competent and thus able to raise its load, forming at first a simple overturned anticline. Under increasing horizontal stress shearing planes developed on 45-degree angles and thrust faulting took place, causing the entire mountain block to be pushed up and moved northward for several miles. There has been more than sufficient vertical rise acting in combination with the horizontal movement to account for the present elevation of the range, and the geological evidence shows that this thrusting is as great or greater than the 3.98 miles credited to isostasy. The unloading and weakening of this arch by erosion may have aided and localized the faulting, but the fact that the fault is a thrust should prove that this later faulting and uplifting is due to compression or horizontal stresses and not to the more vertical action of isostasy.

The writer is not altogether certain as to the exact time of this faulting, or to the entire sequence of events which lead up to the present relief of the Uinta Mountains, but the faulting probably began at or near the culmination of the folding in late Cretaceous or early Tertiary, and may have continued or recurred at intervals throughout the Tertiary until Pliocene or later. The present mountains, he believes, are largely due to the compressive forces causing

a continuance or recurrence of movement on this inclined fault plane. Support for this belief is shown by the fact that the points of greater elevation in the Uintas are opposite points where the fault plane is steepest and the range is lower opposite the great overthrusts. The word "relief" has been used thus far as differential relief or elevation in regard to the surrounding plain, disregarding the epeirogenic uplift. This uplift and differential erosion are both factors to be taken into consideration in the actual elevation of these mountains above present sea level. The epeirogenic uplift is thought to have been in the nature of an up-warping which raised the Rocky Mountain and Basin regions bodily a mile or more in or about Pliocene time. Differential erosion is a geological agent which, of course, played a part in carving out the topography of the Uintas.

The folding and faulting of the Uinta Mountains has occasioned some crustal shortening. The writer knows of no previous attempt having been made to work out the amount of this shortening, so a rough estimate has been made by measurements taken on a scale drawing of the reconstructed fold. Roughly it amounts to 6 miles for the folding and an average of 2 miles for the thrust faulting. On an approximate basis it shows that this section of crust has been compressed into about three-fourths to two-thirds of its original width. Dr. Chamberlin says:[1]

> The inadequacy of isostatic adjustment to occasion crustal shortening on the scale observed is, perhaps, the greatest shortcoming of the isostatic theory of earth deformation. It is rather generally recognized by geologists that the great tangential thrusting exhibited in the folding and faulting of so many portions of the lithosphere cannot be explained by simple vertical isostatic movements.

This crustal shortening certainly is not due to isostasy, and most of the diastrophic events or movements shown by this range tend to give crustal shortening. The present structure shows that the later movements, thrust faulting predominating, are also the result of the same forces. As the vertical action of isostasy is contrary to the observed geologic evidence and as all of these movements may be explained more readily by compression, the writer believes that compression has been the dominant factor in the diastrophism throughout the entire history of the Uinta Mountains.

[1] *Op. cit.*, p. 9.

REVIEWS

Bruch- und Fliess-Formen der Technischen Mechanik und ihre Anwendung auf Geologie und Bergbau. Band II. *Scher-Form.* Band III. *Zerreiss-Form.* By ERICH SEIDL. Berlin: V D I, Verlag, 1930.

The title of the work, of which the two small volumes listed above are the first published thus far, states accurately its aim and scope. Volume II covers twenty-two pages; Volume III, eighty-eight pages. Over one-half of these pages are occupied by illustrations.

The *aim* is to describe and figure the fracture patterns and other changes in form and texture which occur in bodies of diverse shapes and materials when they undergo permanent deformation in the course of industrial testing. A number of characteristic patterns are described and analyzed qualitatively in terms of stresses and resulting strains. An effort is made to recognize in the patterns described such features as are diagnostic of specific conditions of stress. This constitutes the first part of each volume.

In the second part, geologic structures are figured which resemble in ground plan and in section the patterns recognized in the first part as typical of specific conditions of stress. In this way many geologic structures are given a mechanical interpretation, in some cases along lines quite different from those so far suggested.

The *scope* of the whole work is shown by the titles of the separate volumes.

BAND

I. *Theoretische Begründung* (theory).
II. *Scher-Form* (patterns of deformation produced by shearing).
III. *Zerreiss-Form* (patterns of deformation produced by tearing).
IV. *Zerdrück-Form* (patterns of deformation produced by failure under compression).
V. *Biege- und Knick-Form* (*Spröde Falten*) (patterns of deformation produced by bending and kinking, that is, folding of relatively brittle materials).
VI. *Strömungs-Form* (*bildsame Falten*) (patterns of deformation produced by flowage, that is, folding of non-brittle materials).

The geologist, who wishes to pass beyond the mere description of geological structures to attempt an understanding of the mechanics of their

formation, should begin with the analysis of deformations obtained under known conditions of stress. The recent great advances in the knowledge of the physics of materials and in methods of testing has created a wealth of observational materials invaluable for this purpose. Unfortunately, much of this remains unpublished or is printed in publications with which the geologist is not familiar. In collecting and putting at the geologist's disposal some of this material, Seidl has called the attention of the structural geologist to this store of experimental observations thus far neglected. This double service deserves our gratitude.

Volume III, on the forms of tearing, is especially interesting as geologists in the past have limited their own experiments almost exclusively to deformation produced by compression. In the first part of the volume the permanent deformation under tensional stress is described for such bodies as rods of steel, lead, and bronze; rods consisting of single crystals of zinc, aluminum, and rock salt. The influence of various factors is discussed, factors such as rate of application of stress, temperature, structure, and texture of materials (especially of layers of different physical properties), and of the form of the bodies undergoing deformation. Where a less plastic layer lies between more plastic materials, tension causes the former to tear, while the latter flow into the gap thus created. The former are called "*Zerreiss-schichten*," the latter, "*Ausgleichsmassen*." The resulting pattern, in section and in ground plan, constitutes, in Seidl's opinion, the prime criterion for the recognition of structures produced by tearing.

In the geologic part, illustrations are given first of small-scale fractures indicating tearing, such as Heim's well-known pictures of stretched belemnites and echinoderm fragments from Jurassic limestones of the Alps; lenticular masses of limestone between shaly beds in highly deformed strata, etc. Structures of larger scale, interpreted by Seidl as due to tearing, include, e.g., the salt zone of the Upper Allerthal in the salt basin of Central Germany; structures in the coal basins of Lower Silesia and of Forest of Dean, England; dolomite beds in Upper Silesia; anhydrite and rock salt beds in Southern Germany; the Comstock lode; the Triassic basin of Connecticut; the Baberton region of Swaziland; the northern limestone belt of the Eastern Alps; the Red Sea; the cryptovolcanic structure of Steinheim, Germany, and the Ries basin nearby. Although not all the interpretations given are convincing, especially since the scope of the book does not permit discussion of structural details, all deserve careful attention and evaluation.

Volume II, devoted to the forms of deformation produced by shearing,

offers similarly valuable, though more generally known, materials. In view of the importance of shearing in nature, strongly emphasized by Seidl himself, one might consider the space given to it inadequate (twenty-two pages only). But as shearing enters so largely into the deformation of bodies under compression and under bending, much additional material may be expected in the corresponding volumes of this valuable work.

W. H. BUCHER

Ferns, Fossils, and Fuel. The Story of Plant Life on Earth. By ADOLF CARL NOÉ. Chicago: Thomas S. Rockwell Company, 1931. Pp. 128; pls. 8. $1.25.

This book is a semi-popular account of the development of plants throughout geologic history, together with a discussion of the relationship of plants and plant fossils to coal and petroleum geology. It is well worth the while of both the layman and the professional geologist.

Special attention has been given to the Carboniferous floras, and the flowering plants of the Upper Cretaceous and Tertiary. The Mendelian laws and their application are discussed, together with the possibility of parallel evolution in plant history. In chapter iv appears an excellent correlation table of the Pennsylvanian of North America with that of Europe and Siberia.

By an unfortunate error the first sentence in the last paragraph on page 54 reads, "The trilobites became extinct by the end of the Silurian period."

H. W. STRALEY, III

The Japanese Earthquake of 1923. By CHARLES DAVISON. London: Thomas Murby & Co., 1931. Pp. 125; figs. 32; pls. 6. 7*s* 6*d*.

The author gives an interesting account of this great disaster in a clear and non-technical manner. The historical and economic aspects are discussed in the first half, and the geological in the last half. The seismic history and structure of this region are given as a background for the event; the nature of the quake itself is discussed in some detail, followed by a good account of the results. The great discrepancy between the displacements on land and under the water should be noted—these would seem to be due to inaccuracy in soundings.

J. T. McC.